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THE BRAZEAU RIVER VALLEY THROUGH THE FOOTHILLS OF ALBERTA.

Photograph taken by Mr. J D Boothman from an altitude of 10.7 km A.S.L. over the Bighorn Range.

THE UNIVERSITY OF ALBERTA

THE GLACIAL GEOMORPHOLOGY OF THE BRAZEAU RIVER VALLEY; FOOTHILLS OF ALBERTA

by

© DONALD RAYMOND KVILL

A THESIS

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ABSTRACT

It is generally accepted that the distal margins of a glacier furnish the greatest amount of information concerning the size and disposition of the ice body. In the Foothills of Alberta the glacial deposits and landform suites indicate that not only is this an area where glaciers of both Laurentide and Cordilleran origin terminated during Pleistocene advances, but at least once the two ice masses coalesced and flowed southeastward parallel to the strike of the Foothills. The glacial event during which this coalescence occurred is controversial but many researchers (summarized in Rutter and Schweger, 1980) have expressed the opinion that this happened during Early or Pre-Wisconsin time. This study suggests an alternate model of glaciation in which coalescence occurred during the Late Wisconsin advance. Support for this model is provided by a systematic description and interpretation of the glacial landforms in one area of the Foothills where the Quaternary geology has only been described in a very general study (Reimchen and Bayrock, 1977). It was thought that by directing the primary focus of the study toward geomorphological evidence, rather than glacial stratigraphy, the results could provide supplementary evidence which might assist in the development of a regional model of glaciation. Inconsistencies between the geomorphic and stratigraphic evidence will provide obvious topics for future research.

The interpretation of the geomorphic evidence suggests that Cordilleran ice issued through the Brazeau River valley at least as far as the eastern side of the Foothills. Eskers found within the Foothills indicate that the glaciers were not restricted to the major valley systems and that the ice within these valleys became locally stagnant during deglaciation. Scattered erratics and erosional evidence suggest that the upper ice limit in the Foothills was at an elevation of about 2000 metres. A dated core of lake sediment from Fairfax Lake, located at the eastern edge of the Foothills, indicates that this area was exposed from beneath glacier ice approximately 12,000 years ago or earlier. Local ice-flow indicators suggest that the last glacier to traverse the eastern region flowed from the northwest, probably from the Athabasca Valley. Perched glaciolacustrine sediments along the Brazeau River valley (near the eastern edge of the Foothills), the pattern of meltwater channels near this area, the orientation of ice-moulded landforms produced by the ice to the east of the Foothills, and a lateral moraine within the

easternmost Foothills valley, indicate the contemporaneity of the two ice bodies. Regional topographic constraints, augmented by the interference of the two ice masses, reduced the activity of the ice along the eastern margin of the Foothills during deglaciation. This is evidenced by esker complexes and glaciofluvial deposits on the Western Alberta Plains adjacent to the Foothills. The failure of the ice-moulded landforms, created by the ice flowing from the northwest, to deflect around positive relief features such as the Tertiary Uplands and the eastern edge of the Foothills, indicates that some obstruction was preventing the Cordilleran ice from draining toward the east as local topography would otherwise dictate. Extensive glaciofluvial and glaciolacustrine deposits, large meltwater channels oriented toward the southeast, and tills of mixed Laurentide and Cordilleran origin (mapped in the vicinity of the Brazeau Reservoir east of the study area) indicate that Laurentide and Cordilleran glaciers coalesced in this area. Deflection of the Cordilleran ice by the Laurentide Lobe therefore provides a plausible explanation for the observed ice-flow indicators.

The observed landform associations can all be explained in terms of a single glacial event. No evidence was found of a subsequent, less-extensive advance except in the high alpine areas. It is therefore concluded that the geomorphic evidence along the Brazeau River valley does not support the theory of the Foothills having escaped glaciation throughout the Late Wisconsin glacial event.

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1. 1. INTRODUCTION, STUDY AREA AND METHODOLOGY

1.1 INTRODUCTION

The glacial history of the Rocky Mountain Foothills has proved to be perplexing as well as controversial. One of the problems inhibiting the development of a reasonably precise glacial chronology is the fact that many areas have not been mapped or described at more than a reconnaissance level. The Brazeau River valley is one area where very little geomorphic and stratigraphic work has been done. Therefore, one of the principal functions of this study was to map and describe the glacial landforms along the Brazeau River valley from the First Range of the Rocky Mountains to the Brazeau Reservoir (see Fig. 1.1). From the data presented in this study an interpretation is developed to suggest a model of ice behavior during the last major glacial event. In addition, an interpretation is offered as to the nature of the ice lobes from various sources. It is hoped that this will shed light on the problem of an ice-free corridor along the front of the Rocky Mountain Foothills during late Pleistocene time.

Problems remain but it is expected that the proposed glacial sequence will stimulate additional research in this area. Closely related themes which could be investigated further include the geomorphic mapping of the Foothills, between the area mapped in this study and the Athabasca River valley to the north, and additional palynological work on the lake cores from the region.

1.2 STUDY AREA

1.2.1 LOCATION OF STUDY AREA:

The study area was defined so as to transect the Rocky Mountain Foothills, as depicted in the Atlas of Canada (1957) and the Atlas of Alberta (1969), and extend into the Interior Plains as far as the western boundary of the area depicted on the NTS 83B mapsheet. The Quaternary geology of that mapsheet area has been mapped by Boydell (1972). Ideally, the present study should have extended to the valley heads within the Rocky Mountains, but extensive studies of the related aerial photographs, and aerial

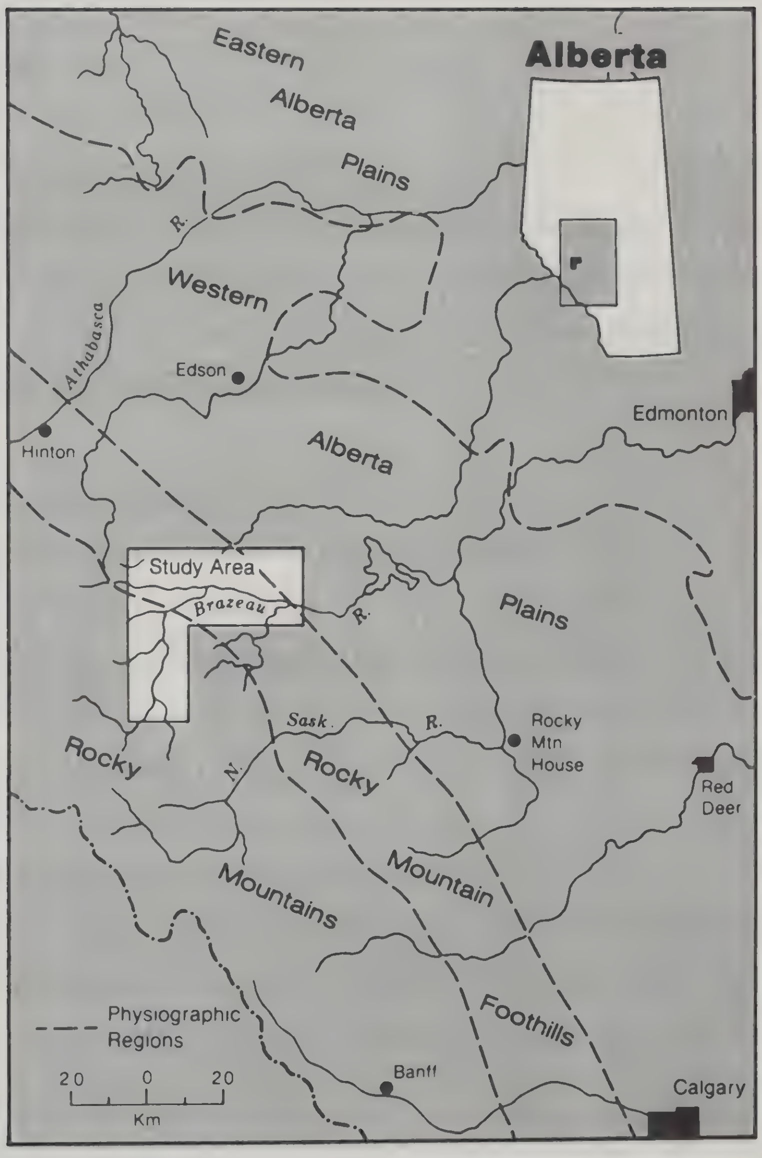


Fig. I.I Study Area Location Map

reconnaissance, persuaded the author to restrict the study to the area downvalley from the mountain front. The mapping was therefore restricted to an area embracing the following NTS 1:50,000 mapsheets:

(1) George Creek 83C 10/W

(2) Grave Flats 83C 15/W

(3) Pembina Forks 83C 15/E

(4) Blackstone River 83C 16.

The interpretations and conclusions, however, are based on field and aerial investigations of a larger area which includes the upper Brazeau River valley depicted on parts of NTS 83C 6 and 83C 7 and the contact zone between Cordilleran and Laurentide ice which lies within the 83B 13 mapsheet area.

1.2.2 JUSTIFICATION OF STUDY:

The study was developed for the following reasons:

(1) The geomorphology of the area has not been previously mapped. A study conducted by Reimchen and Bayrock (1977) covered part of the area but did not extend east beyond the edge of the Foothills. No preliminary soils maps have been prepared for the area and the geologic maps, except for private maps produced and owned by mining companies, generally do not include the Quaternary deposits (e.g. Allan and Rutherford, 1924; Holter and McLaws, 1974).

(2) The portion of the Athabasca Valley depicted on the NTS 83F mapsheet has been mapped and described by Roed (1968, 1975). His study area lies almost immediately north of the area of the present study. Boydell (1972;1978) mapped the NTS 83B mapsheet area which lies adjacent to the eastern margin of the present study area. Therefore, the present study will serve to link these earlier reports and provide additional, regional, interpretative material.

(3) The coalescence of Cordilleran and Laurentide ice during the Late Wisconsin glaciation has been most clearly documented by Roed (1968, 1975) for the Hinton–Edson area. Boydell (1972; 1978) suggested that coalescence occurred in the Rocky Mountain House area during the Late Wisconsin (Jackfish Creek) advance (Boydell, 1978, p.24). Further south, along the Foothills Belt, the geomorphic

evidence suggested that the Cordilleran and Laurentide advances were nonsynchronous and that an ice-free corridor existed for at least the last 50,000 years (see Reeves, 1973, and Rutter and Schweger, 1980, for a more complete discussion).

The ice-free corridor concept is frequently cited in anthropological literature as a possible migration route for early man into North America. This concept was the principal topic of the 1978 meeting of the American Quaternary Association held in Edmonton, Alberta (Rutter and Schweger, 1980). However, the paucity of radiometric dates and critical geomorphic evidence has allowed the controversy to continue as to when, where, and for how long the ice-free corridor was open (Mathews, 1980; Rutter and Schweger, 1980). This study attempts to provide geomorphic evidence and radiometric dates that will help to resolve some of this controversy.

(4) The study area lies within a portion of Alberta which is currently undergoing considerable development for the production of fossil fuels and timber. The Luscar Coal Company has developed a large open-pit mine for the extraction of bituminous coal adjacent to the north side of the study area. Many of the major petroleum companies are currently exploring the area for oil and natural gas. Commercial Timber Permits have been issued along the Pembina River and the entire area has been identified by the provincial government as a Proposed Timber Development Area. A map of the glacial deposits which mantle much of the area will provide useful information for the purposes of terrain analysis, location of industrial gravel deposits and slope stability studies.

1.2.3 PHYSIOGRAPHY AND GEOLOGY:

The headwaters of the Brazeau River form within the Front Ranges of the Rocky Mountains. The south fork of the river originates on the eastern side of Nigel Pass which occupies the interfluve between the Brazeau and North Saskatchewan drainage basins. The north fork heads at the Brazeau Icefield complex on the southeast flank of Mount Brazeau. Both forks of the river occupy valleys which show clear evidence of glacial erosion. Downstream from the confluence of the two branches the glaciated valley

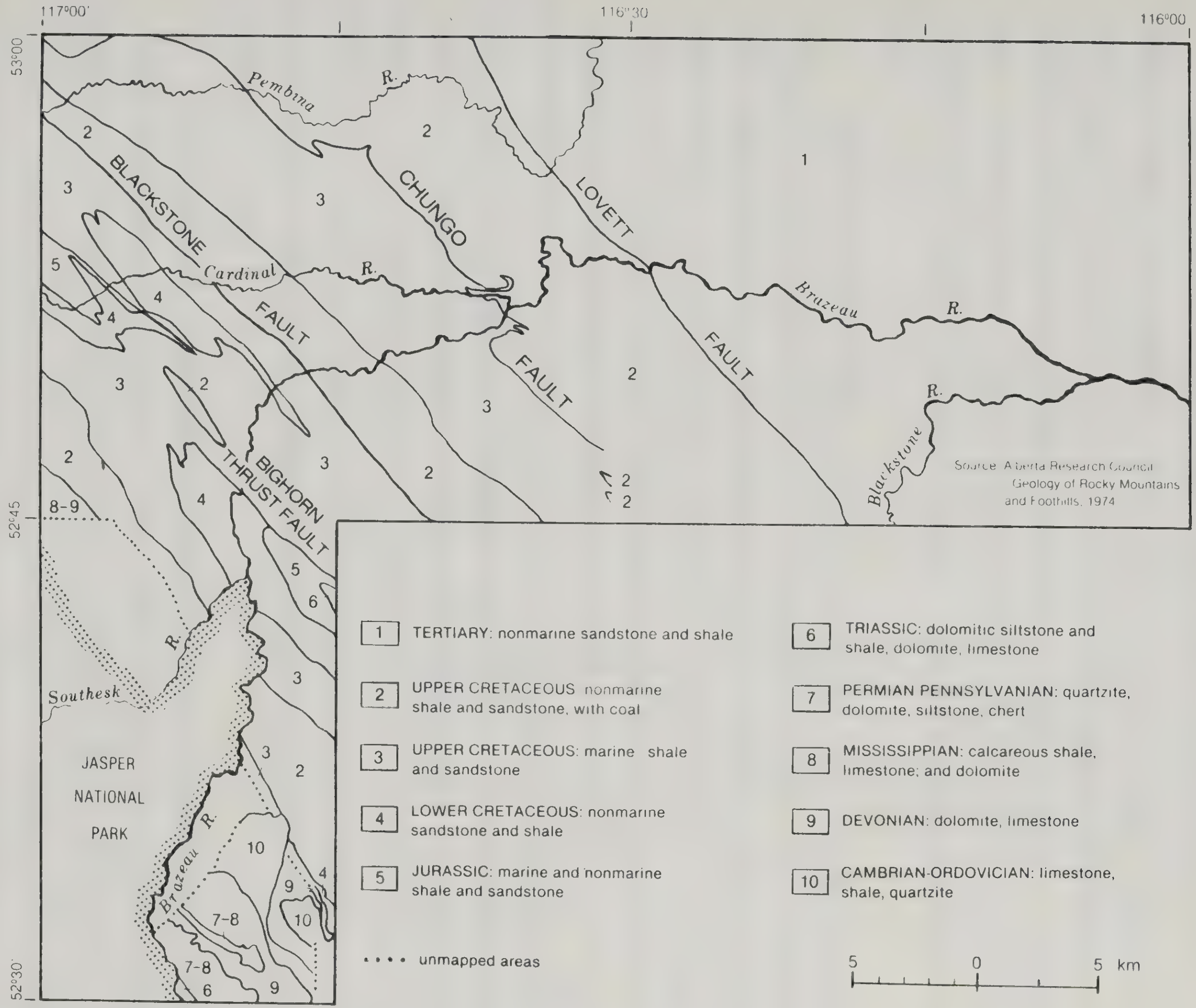


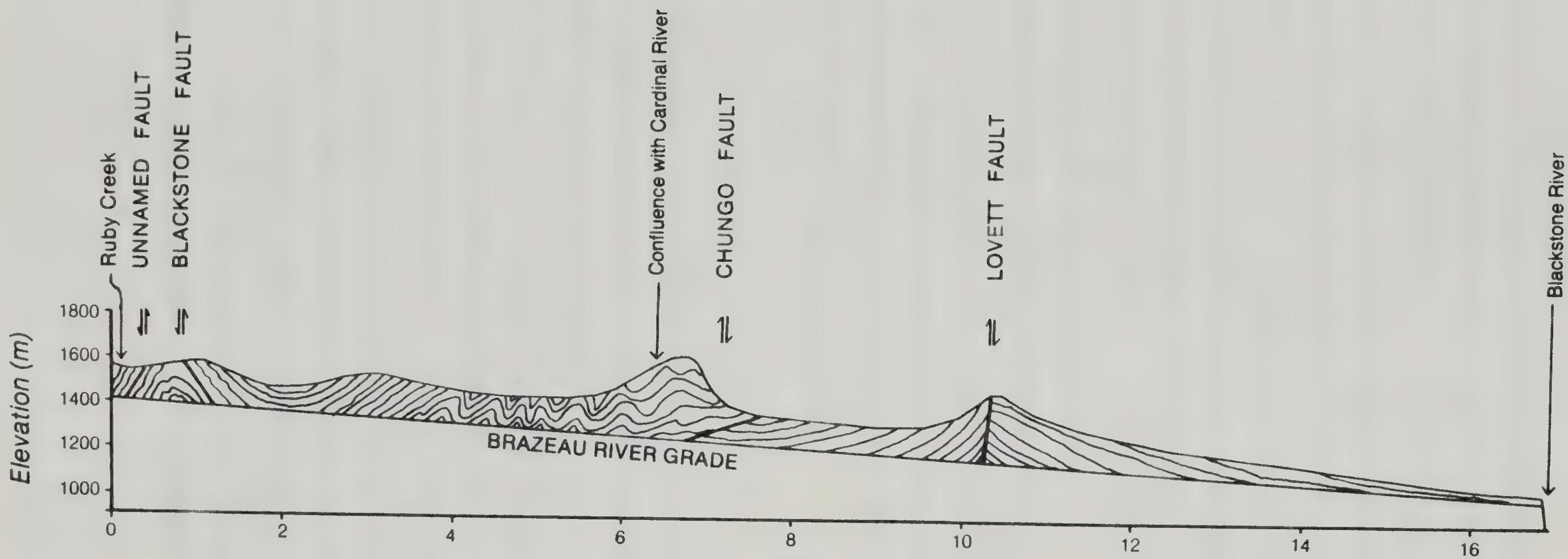
Fig. 1.2 Bedrock Geology

widens and deepens as it passes through the eastern Front Ranges. Local relief along the valley commonly exceeds 1400 m. The structure of the Front Ranges consists of a relatively simple series of thrust faults (see Figs. 1.2 and 1.3).

The oldest rocks, of Precambrian age, are found along the south fork of the Brazeau system. The sedimentary units become generally younger eastward to the eastern edge of the Foothills where Tertiary material underlies the glacial deposits. The Paleozoic and Early Mesozoic sediments are dominantly marine carbonates, quartzites and shales whereas the more recent units are comprised of marine and non-marine sediments which include carbonates, sandstones, shales and coal (Holter and McLaws, 1974).

Upstream from the mouth of Job Creek the Brazeau River is incised into bedrock. Downstream from the confluence with Job Creek the river has incised through thick glacial deposits although bedrock commonly outcrops within the main channel. A sinuous channel pattern is prevalent within the mountains but a braided reach exists through the eastern Front Ranges. The braided reach occurs where the river has incised into the unconsolidated glacial deposits. Good sections of Quaternary deposits, up to 40 m thick, are exposed on either side of the valley above the confluence with the Southesk River. Downstream from the mouth of the Southesk River the Quaternary sediments become thinner and the underlying bedrock rises along the upthrust limb of the Bighorn Thrust Fault. For the next 25 km downriver the valley transects a series of foothill ridges between the Bighorn and Chungo Faults. The valley in this section is deeply incised into bedrock with only a thin veneer of Quaternary deposits exposed along the upper portion of the canyon walls.

After the Brazeau River crosses the Chungo Fault (manifested by the Cardinal Hills) it flows across Upper Cretaceous / Early Tertiary Uplands which are the western extension of the Western Alberta Plains. The topography no longer reflects the strong northwest – southeast orientation of the Cordilleran Ranges but takes on an irregular, rolling countenance with a slight east – west trend observable in the larger features. Relative relief in this area is commonly in the order of 100 m. The river flows within a 40 m deep canyon with bedrock walls mantled by 3 to 5 m of unconsolidated glacial sediments. It is not known how deep the unconsolidated mantle is over the uplands on



Source: Allan and Rutherford, 1924

Fig. 1.3 Structural Geology along the Brazeau Valley

either side of the valley but roadcuts and seismic line cuts rarely extend through the drift and scattered well-log data suggest that 2 – 15 m is a common thickness range.

The morphology of the Upland areas, their position relative to the major valley corridors through the Front Ranges, and the Tertiary age of the bedrock beneath the drift in the area indicate that sediments from the Cordillera were being deposited over parts of the study area prior to the onset of glaciation. Deposits of Pleistocene and Holocene age mantle the consolidated deposits of Tertiary and Cretaceous age.

The area surrounding the fan-shaped Tertiary Uplands is low and comparatively level, with stream channels providing most of the local relief. Organic deposits are common and elevations are generally below 1200 metres.

After transecting the ridge system which marks the Lovett Fault, the Brazeau Valley enters the lower plains area south of the Tertiary Uplands. The valley morphology abruptly changes from a steep-walled bedrock canyon to a broad valley flanked by multiple terraces. The river adopts a braided channel configuration which persists beyond the eastern side of the study area to the Brazeau Reservoir.

1.2.4 DRAINAGE:

The study area includes the drainage divide between the Arctic and the Hudson Bay drainage basins. The actual divide follows a sinuous course between the Brazeau and Pembina river systems. The Brazeau River and all of its tributaries drain into the North Saskatchewan system whereas the Pembina River is within the Mackenzie River drainage basin. At the eastern edge of the Foothills, these two rivers are separated by less than 3.5 km. A topographic depression occupied by a small, unnamed tributary to the Brazeau River can be traced from the west limb of the Big Bend in the Brazeau River immediately to the east of the Foothills, to the east limb of the Pembina River, which also makes a major bend in this region (see Landforms Map). It is highly probable that the events of the Pleistocene diverted flow from the Pembina system to the Brazeau drainage basin (and consequently the North Saskatchewan system) from the drainage network which now feeds into the Beaufort Sea.

1.2.5 VEGETATION:

Except for floodplains, some terraces, bogs and a few small prairies between the First Range and the Foothills the study area is dominated by arboreal vegetation. The lower surfaces on the east side of the study area are classified in the Canada Land Inventory as forming a subalpine ecosystem. This grades westward into a true alpine environment along the higher foothills and the mountains of the First Range (Environment Canada, 1974) Lodgepole pine (*Pinus contorta*) is the dominant tree species but locally aspen poplar *(Populus tremuloides)*, white spruce (*Picea glauca*), black spruce *(Picea mariana)* and balsam poplar (*Populus balsamifera*) form pure stands. In mixed stands, tamarack (*Larix laricina*), Engelmann spruce (*Picea engelmannii*) and balsam fir (*Abies lasiocarpa*) can be found. Willow (*Salix spp.*) and alder (*Alnus spp.*) are frequent members of riparian and fen communities.

Except for areas which are heavily forested, a vigorous shrub component is found in the plant communities. Heavy winter snow appears to favor the development of shrub communities. It overtops the developing shrubs protecting them from predation and wind and snow damage during the winter. Buffaloberry (*Shepherdia canadensis*), shrubby cinquefoil (*Potentilla fruticosa*), silverberry (*Elaecagnus commutata*) and bog birch (*Betula glandulosa*) are common in open meadows and on grassy slopes. Juniper *(Juniperus spp.)* is most commonly found on the more xeric slopes and Labrador tea (*Ledum groenlandicum*) is common in hygric sites.

Abundant forbs occupy areas where soil drainage is adequate. Disturbed areas such as cut-lines, abandoned campsites and petroleum leases, as well as naturally disturbed areas such as forest fire scars, mass movement features and game trails, are invariably colonized by a rich assemblage of forbs and grasses. Dandelions (*Taraxacum officinale*), fireweed (*Epilobium angustifolium*), sow thistle (*Sonchus arvensis*) and avens (*Dryas spp.*) are common pioneer species in disturbed areas. Strawberry (*Fragaria glauca*), arnica (*Arnica spp.*), milk-vetch (*Astragolus spp.*), Loco weed (*Oxytropis splendens*), twinflower (*Linaea borealis*) and bunchberry (*Cornus canadensis*) are common in mesic sites with adequate illumination (Cormack, 1977).

Within the Foothills, particularly adjacent to the mountains, there are numerous areas where grasses provide the dominant cover. Many of these prairies are not easily

explained as a seasonal moisture deficit is not the obvious limiting factor that precludes arboreal vegetation. Perhaps a seasonally high water table or desiccating winter winds may be responsible for furnishing grasses with the necessary competitive advantage in these locations.

Numerous species of graminoids can be identified, the most visible of which include wheatgrass (*Agropyron spp.*), Brome grass (*Bromus spp.*), Reed grass (*Calmagrostis spp.*) and fescue (*Festuca spp.*). In hygric environments the graminaea are replaced by sedges (*Carex spp.)* (Environment Canada, 1974).

1.2.6 CLIMATE:

The paucity of year-round recording stations, plus the location of all long-term, year-round stations in valley bottoms, has precluded detailed studies of the local climate along the eastern slopes of the Rocky Mountains. From the data available Longley (1970) classified the entire area as a Sub-Arctic or Cold Snowy Forest climate using the Koppen System of classifying climates. The Canada Land Inventory studies (Environment Canada, 1974) describe the region as having a cold, sub-humid climate with a short growing season. The frost-free period in the Foothills is usually less than 60 days and this decreases toward the mountain front.

Climatically the northern Foothills differ considerably from those in southern Alberta because of a relatively reduced influence of airstreams from the Pacific. Reinelt (1970) and Powell (1977) have indicated that the rainshadow effect is only significant in these areas during the winter months and even then the location of the Arctic Front is usually well to the south of the study area. Thus the path of cyclonic disturbances from the Pacific is usually well to the south of this area and stable, dry conditions prevail. Both studies indicate that summer precipitation patterns reflect orographic (up-slope) influences plus heavy convective precipitation in the early summer. Differences in aspect, surface texture, moisture conditions and albedo, resulting from the topographic diversity of the area, augment the atmospheric instability of early summer (Reinelt, 1970).

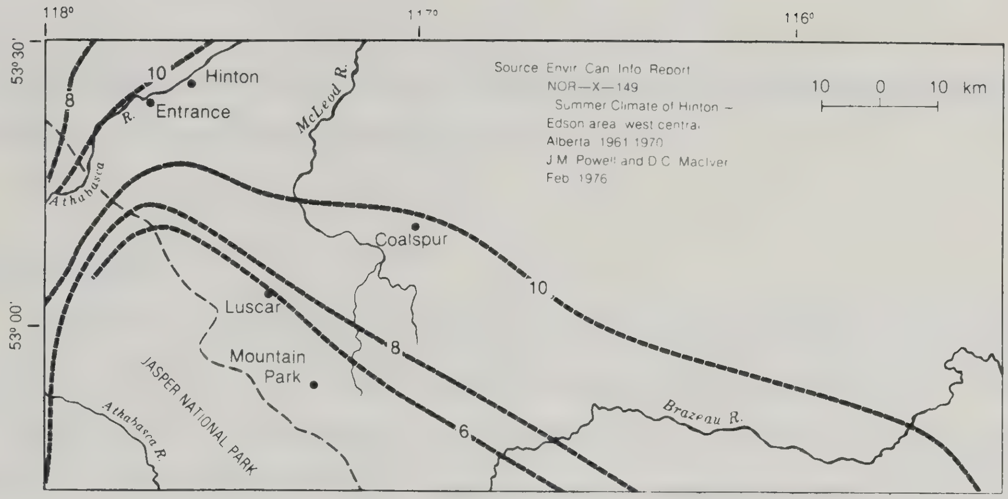
The significance of convectional and orographic precipitation along the eastern slopes is illustrated by the fact that the Clearwater Forest (which includes most of the study area) experiences an average of 20 thunderstorm days during the growing season,

compared to less than 10 at stations within the mountains such as Jasper (Powell, 1977, p.4). Rocky Mountain House experiences an average of 125 days per year with measurable precipitation whereas Jasper only records an average of 89. Reinelt (1970) calculated that approximately 18 per cent of the precipitation which falls within the study area is stimulated by orographic effects. Sixty-eight per cent of the precipitation falls as rain (Powell, 1977, p.6) and there is no period of the year in which a moisture deficiency exists (Powell and MacIver, 1976, Fig. 16). Snow has been recorded in every month of the year.

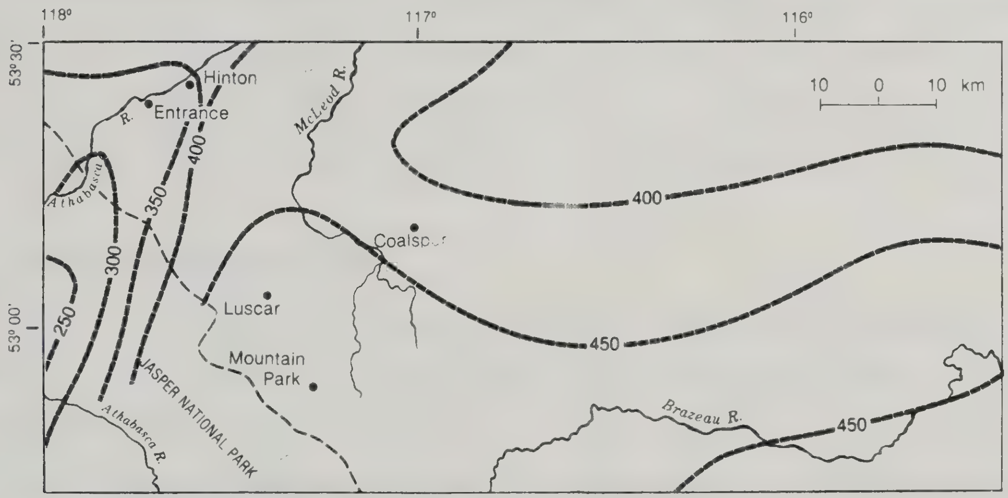
Figure 1.4 illustrates the mean May to September temperatures and the total number of days during this time period where the mean daily temperature exceeds -2.2 degrees Celcius This threshold temperature is used, instead of zero, because it is considered to be a killing frost for most sensitive indigenous vegetation (Powell and MacIver, 1976, p. 15). The isotherm pattern depicted by Powell and MacIver (1976) reflects the influence of topography on the temperature pattern. The high plains or uplands to the east of the foothills are relatively warm but a steep temperature gradient exists between them, the Foothills and the higher First Range. The Athabasca River valley is considerably warmer than the adjacent mountain and Foothills area. It is important to note that nearly all of the study area experiences 13 or more days during each growing season when the mean daily temperature fails to exceed the -2.2 degrees Celcius. threshold. The western portion of the region is even colder, with at least 28 days during the May - September period when the mean daily temperature fails to exceed the threshold value.

The precipitation map (Fig. 1.4) portrays isohyets which are not well aligned with the physiographic structure. The only year-round, long-term, recording stations are located at Ranger Stations which are invariably at low elevations within the major valleys. Summer data are recorded at all forest lookout stations which are located only on high vantage points. It is therefore probable that the irregularities in the isohyet map are artefacts of the data sites rather than an accurate reflection of the precipitation pattern.

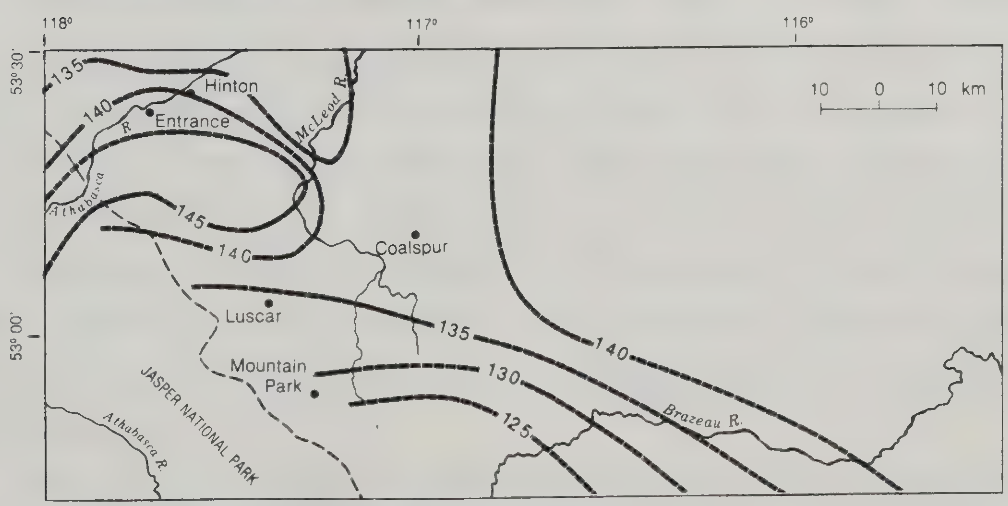
A zone of relatively high precipitation extends across the northern end of the study area. This zone generally lies along an upland area that is only about 305 m (1,000 ft) or less above the surrounding lowlands. The intensification of this pattern in the



Mean seasonal temperature (°C) May to September, 1961-1970.



Total seasonal precipitation (mm), 1961-1970.



Total seasonal number of days, May to September, with mean temperature >-2.2°, 1961-1970.

Fig. 1.4 Climatic Maps

months of June and July, when most of the precipitation falls, underscores the role of orographic and convectional rainfall in this area.

1.3 METHODOLOGY

Historically, the glacial chronologies of the Eastern Slopes of the Rocky Mountains have been studied mainly by glacial stratigraphers (Rutter, 1965; Roed, 1968, 1975; Boydell, 1972, 1978; Reimchen and Bayrock, 1977; and others). In these studies, helicopters, truck-mounted drills and large field parties were often used to gather data. The limited personnel, equipment and access precluded a similar approach in this study. A methodology had to be devised which could be conducted by a single researcher over a study period of two summers. The methodology thus consisted of mapping the landforms from a combined airphoto, aerial and ground investigation of the study area.

1.3.1 AIRPHOTO INTERPRETATION

Before the fieldwork commenced an airphoto interpretation of the area was carried out using primarily the 1:40,000 scale aerial photographs available from the University of Alberta Map Collection. A complete collection of the study area photographs, taken in 1950, was used for the actual mapping. The 1950 series is the only complete coverage of this portion of the eastern slopes of the Rocky Mountains. The quality of the photographs is adequate for geomorphic interpretations partly because the forest cover is less extensive than at present because of a higher incidence of forest fires in the two decades before 1950 (Ross, 1974). Supplementary information was obtained from near infra-red airphotos (scale, 1:21,120), LANDSAT images and larger-scale panchromatic aerial photographs. The initial mapping was carried out directly onto the airphotos using water-soluble ink. In this way landform boundaries could be accurately mapped and alterations could be more expediently made than with the use of acetate overlays.

The original objective was to identify and map all of the landforms from aerial photographs and to field-verify the mapping during the fieldwork phase of the study. This proved to be impractical. During the initial airphoto interpretation areas were identified where the characteristics of the landforms were uncertain. The questionable

areas were marked on a topographic map and aerial reconnaissance and ground observations were conducted to provide essential information on the sedimentological properties and explicit morphology of the features concerned. In some instances repeated investigations of particular areas were required before the nature of the landforms could be determined. For this reason a mirror stereoscope was retained at the base camp and a pocket stereoscope, plus the airphotos of the area, were carried during all field traverses.

At the end of the second field season a preliminary draft of the landforms map was prepared using NTS 1:50,000 topographic maps as a base. The data were transferred from the airphotos to the maps using a transfer scope. Inaccuracies were found because only the provisional maps, based on the topographic and survey data obtained by the geological surveys during the 1920's and 1930's, were available during the fieldwork phase of the study. The landforms were depicted as accurately as possible on these maps and the information was directly transposed to the final map which is based on more recent topographic information. The initial airphoto analysis was conducted using a Keuffel and Esser mirror stereoscope and 1:40,000 scale black and white panchromatic aerial photographs. The landforms were identified on the basis of morphology, drainage patterns, evidence of erosion, and the relationships between the landforms. The vertical as well as planimetric relationships between the landforms was noted. The surficial deposits were determined by the use of indirect indicators such as drainage patterns and density, tone and texture, and the vegetation cover (for a more complete discussion see Colwell, 1960; Lattman and Ray, 1965; or Estes and Senger, 1974). It was found that in this area, deciduous trees, especially aspen poplar (*Populus tremuloides*) and white birch (*Betula papyrifera*) were good indicators of well drained soil conditions, hence the parent material of the soil was inferred to be usually coarse textured. Black spruce (*Picea mariana*) and tamarack (*Larix laricina*) were good indicators of hygric sites and lodgepole pine (*Pinus contorta*) was most commonly found on mesic sites. Sedges and hygrophytic shrubs (eg. willow and birch) were indicative of saturated soil conditions. In the Foothills, vegetation should be used with a degree of caution as an indicator of soil drainage conditions. Other factors, such as variability of microclimate, the depth of the unconsolidated deposits, angle and aspect of slope, plus

fire history may mask the differences caused by texture of the parent material.

The final airphototo investigation was conducted using an Old Delft scanning stereoscope. Most of this study was conducted using the standard 1.5X magnification but the 4.5X option could be selected to enlarge the image for critical areas. The excellent optics and the scanning feature of this instrument are desirable for detailed geomorphic work.

1.3.2 AERIAL RECONNAISSANCE:

The use of a light aircraft was essential for this study. Aerial photographs furnish three-dimensional views of the terrain but the fixed scale and the single perspective limit the information which can be thus gathered. It has been the author's experience that, for heavily forested environments such as the study area, an airphoto interpretation alone provides only a very crude reflection of the actual landform assemblages and related surficial deposits. An aircraft enabled the observer to change the perspective and scale of the terrain features, thereby permitting a more thorough investigation of the area. Moderate strength binoculars with a wide field of view (i.e. 7X50) were useful for examining exposures which are poorly accessible on the ground. It should be noted that this is also an excellent method of inducing motion sickness!

Because aircraft are useful tools for earth scientists, but have not been commonly utilized for studies such as this, some pertinent observations are listed below:

(1) Whenever possible have a pilot or second pilot who is concerned only with the aircraft during field reconnaissance. It is much too easy for a pilot/researcher to become engrossed in studying the features on the ground to the point that he/she becomes one. In areas of high relative relief a division of responsibilities is absolutely essential and a thorough pre-flight discussion between the pilot and the researcher can only increase the efficiency of the project.

(2) When a choice of aircraft types is available, reconnaissance is usually easier from a slower aircraft, particularly if it has large windows or doors which can be removed. However, slower aircraft are generally less comfortable because of their greater sensitivity to turbulence. This results from the decreased wing loading and high-lift characteristics of the airfoils. The rougher ride is partially offset by the fact that

fewer turns and passes are required at slower speeds. If the additional expense is warranted, a helicopter is significantly better than a fixed-wing aircraft for reconnaissance. The visibility is comparatively good and the speed may be decreased without compromising safety.

(3) The additional safety thought, by some, to be provided by a multi-engined aircraft during low-level reconnaissance is largely illusionary. At slow speeds and low altitudes an engine failure on most light multi-engined aircraft, especially if the aircraft is loaded to near maximum gross weight, will produce a critical situation.

(4) Aircraft altimeters are useful for determining the elevation of features where access is difficult. In this study the elevations of cirque thresholds, cols, rock glaciers, breaks-of-slope on valley walls and moraines were determined from aerial observations. Aircraft altimeters are sensitive to atmospheric pressure differences which correspond to altitudinal changes of 20 feet (6.1 m).

Two problems must be overcome in order to obtain accurate elevations from an aircraft altimeter:

a. The Mean Sea Level barometric pressure at the feature to be measured may not be the same as that observed at the nearest Flight Service Station.

b. Difficulty is commonly experienced when attempting to estimate relative heights from an aircraft. In areas of high relative relief it is not uncommon for estimates to be in error by as much as 100 m when attempting to judge accordant heights from an aircraft. The instability inside a moving aircraft precludes the use of conventional leveling instruments but the structure of the aircraft may be used as a sighting device. The turn and bank and artificial horizon instruments indicate when the aircraft is flying straight and level. By then sighting along parts of the aircraft structure, such as wing struts or wing surfaces, the relative elevation between the aircraft and the object to be measured may be determined. It must be kept in mind that aircraft wings are usually not rigged precisely 90 degrees to the vertical axis of the fuselage so it is necessary to compensate for this when estimating accordant elevations. If wing struts are present, small bits of tape or crayon marks placed at eye-level with the observer seated in the aircraft are of great assistance for

estimating relative heights. Alternatively, a thin strip of opaque tape on the window opposite the observer may be used as a horizontal reference.

These techniques are not very precise so the distance between the aircraft and the object being measured must be as short as safety permits. Several readings are required before a reasonably accurate estimate of object elevation can be made. Field checks of this technique were made by determining the height of surveyed mountain peaks. These checks indicated that the method was accurate to within 10 metres.

Frequently relative elevations are more important than absolute elevations for geomorphic purposes. The aircraft altimeter is useful for such determinations because most of the independent variables which influence the accuracy of the instrument may be assumed to be uniform over short distances and time spans (e.g. temperature and barometric pressure). Therefore, the relative heights of features can be measured to a greater degree of accuracy than might be obtained by altimetry on the ground.

(5) Flying at low altitudes in confined valleys should always proceed downvalley and toward the downwind side of the valley where possible. If helicopters are used this is less urgent as turns can be safely executed at slower speeds but even helicopters are vulnerable to the violent downdrafts which may be experienced on the lee-side of mountain valleys.

(6) Flights into mountainous areas should be scheduled for times when meteorological turbulence is minimal. Low windspeeds and stable atmospheric conditions are usually optimal during early morning or late evening hours. Flights at these times are usually preferable to mid-day flights.

(7) Topographic features are visually enhanced by backlighting as the shadows emphasize the relief. This augments the value of morning or evening flights for reconnaissance observations in geomorphic studies.

1.3.3 GROUND INVESTIGATIONS:

Prior to the fieldwork phase of the study a preliminary airphoto interpretation of the area was carried out. This provided the basic geomorphic information for the more detailed ground investigations. A camp was then established at the Brazeau River crossing on the Forestry Trunk Road. This location offered optimal access by vehicles and was previously used for recreational camping so that fire permits were not required and brush removal and creation of waste disposal facilities were unnecessary.

Except for bi-weekly flights to Edmonton for supplies, the camp was continuously occupied from May 28 until September 1, 1977 and from May 23 to September 1, 1978. During the initial fieldwork period extensive use was made of the aircraft for familiarization with the study area and the adjacent areas to the north and south along the mountain front. Access corridors were plotted on the 1:50,000 scale field maps and exposures of the Quaternary deposits were noted and plotted. Parts of the North Saskatchewan River and the Athabasca River valleys were flown and the aerial observations were compared to the published geological reports for these areas.

The Forestry Trunk Road which transects the study area is mapped as an improved road but there are actually no reliable all-weather roads in the study area. Most of the roads are used as access for petroleum exploration, coal exploration and timber exploitation and management. These roads are negotiable by two-wheeled drive vehicles only during the drier periods of the summer or after freeze-up if the snow has been cleared. In late winter tire chains are recommended as icings and ice produced by local snowmelt frequently result in very hazardous grades.

A half-ton truck and a light motorcycle were used for some surface investigations. This combination proved useful as the truck could be employed on the roads and the motorcycle could be transported in the truck to the trails, seismic lines and roads which were not navigable by truck. A light motorcycle was found to have advantages over a heavier machine. Frequently very soft muskegs had to be traversed and the lighter machine would not break the thin vegetative mat on the surface. The lighter machine could also be lifted over obstructions that would have prevented the passage of a larger motorcycle. The potential speed advantage of a more powerful machine would probably not be realized as trail conditions were almost always wet and

very rough. During the fieldwork phase of this study, two new access roads were constructed by companies working on petroleum leases. Frequent observations were made of the fresh exposures created during this construction.

Much of the fieldwork necessitated hiking into areas which were inaccessible by wheeled vehicles. Seismic lines, game trails and horse trails were used where possible but, in many instances, areas were explored by traversing relatively pristine forest. Muskegs, swamps and watercourses greatly inhibited travel in these areas.

One of the first projects undertaken in the field was to assess the applicability of the relative age-dating techniques used by Quaternary researchers to estimate the relative ages of deposits. Characteristics of soil development, leaching of calcium carbonate, weathering rind thicknesses and surface boulder frequency were investigated. However, it was found that the variability of microclimate and parent material was such that any observed differences in soil development or carbonate leaching could be more easily explained in terms of local site conditions than differences in the age of the deposits (see Section 4.2.6). The comparison of weathering rinds was also rejected as a dating device because most of the clasts are carbonates or quartzites. The quartzites are extremely variable in hardness and thus they weather at different rates.

Surficial deposits were bulk sampled only where field descriptions were inadequate to determine distinctions such as texture, lithology or organic content. These samples were later analyzed in the laboratory. Till was the most common material sampled and bulk samples as well as pebble collections were made. Samples of till for a representative transect were collected from the eastern slopes of the First Range to the eastern edge of the study area. Bulk samples of till (approximately 2 - 3 kg) were first removed and then additional samples were removed from the exposure and sieved at the site (through a 1 centimetre-mesh sieve) to separate out approximately 1 kg of pebbles. The pebbles were later classified in the laboratory into major lithologic categories. The locations of the sites sampled are illustrated on Figure 4.1.

Stratigraphic sections were measured and described where natural or man-made exposures revealed a cross-section of the surficial deposits. Generally, thicknesses of the deposits were determined by extending a measuring tape down the face of the exposure and recording distances from the top of the exposure. Slope angles were

recorded so that the thicknesses of the units described could be calculated. Descriptive notes were taken of the sedimentary units and structures observed at each of these exposures. Because the major focus of this study pertains to the area's geomorphology, and not stratigraphy, the sections were not logged in great detail. No litho-stratigraphic units were defined but glacial, glaciofluvial and glaciolacustrine units were described and measured where natural exposures or road cuts provided sections. The principal objective of the limited stratigraphic work was to determine the general sedimentary characteristics of the surficial deposits comprising the local landforms. In addition, a careful search was conducted for nonglacial deposits, weathered surfaces or paleosols separating glacial sediments because these could provide evidence for multiple glacial events in the area.

1.3.4 MAP PRODUCTION

The visual presentation of the interpreted geomorphic data could have been accomplished in a number of ways. Geologic maps usually use color to denote rock types and letters to denote the geologic age and formal stratigraphic unit. Surficial geology maps express the bedrock and/or sediment types as a color with lettered symbols designating genetic class and morphology of the deposits. Screened patterns may be superimposed on the colors to indicate discontinuous, complex or extraneous units (i.e. intermittent bedrock outcrops, intermittently flooded areas etc.). Both of these mapping techniques allow the presentation of more than one symbol on the same unit. This is usually done by placing one symbol as the numerator and the other as the denominator of an algebraic fraction to indicate that the units are superimposed in the manner suggested by the symbols. In some cases numeric qualifiers are included to inform the reader of the relative proportion of one unit to the other. For example 7a/3b would indicate that seven-tenths of the mapped unit consists of parameter **a** and three-tenths of the area has the characteristics of parameter **b**.

Perhaps the most complicated use of compound letters and number symbols is used on the Land Classification and Soils maps. The unit designators used in these maps commonly consist of more than a dozen letter-number combinations superimposed on colored units and overlaid with symbols and patterns.

It is obvious that the number of units which may be included on a map is practically limitless. The problem with complex unit designators is that the visual impact of the map is sacrificed for detailed information. Such mapping techniques should therefore be used where specificity is important to the perceived user. This is quite different from the objectives of the Landforms Map included in this study. The major objective of this map is to present the information in a manner which is visually meaningful to a reader knowlegeable in glacial landform assemblages. Admittedly this is an unusual method of presentation, but it is hoped that the product will generalize and display the information in a manner that provides a visual display of the model of ice disposition discussed in Chapter 5. Color printing is difficult (expensive) when more than three colors (in addition to black and white) are used, so screens and patterns were used to vary the tone of the colors. The colors denote genetic units and the tone (value) of the color was used to portray an impression of topographic position. The darker tones were used on units which occur at higher elevations and lighter tones were assigned to landforms that are found at lower elevations. Obviously this objective can only be met in a general way as many of the landform units occur throughout a range of elevations. The contour lines are the only reliable topographic information.

Screened patterns were required in order to generate all of the nineteen units represented on this map. Again, the selection and arrangement of the patterns were designed so that each would be visually meaningful as well as distinct from adjacent patterns.

The principal justification of electing to use a non-standard approach for this map is that the purpose of the map is distinctly different from that of a surficial geology or a soils map. The objective of the Landforms Map is to illustrate, in three dimensions, the spatial relationships between the genetic units. It is assumed that visual impact and the spatial relationships take precedence over the alternative of providing detailed site information. The cartographers employed during the production of this map advised that patterns and colors are perceptually easier to identify, group and compare than are colors or symbol clusters. In addition, if a designator was included on the units to indicate sedimentary characteristics, the information would have been extrapolated between observation points and in many cases this would have been misleading. It is felt that the

system used displays the level of accuracy that the field evidence can support.

1.3.5 SOIL ANALYSIS:

Most of the soils in the study area have developed on till of Cordilleran origin. Mathews (1980) distinguished between Cordilleran and Rocky Mountain tills on the basis of the presence or absence of lithologies which had been transported from west of the Rocky Mountain Trench. In this study the term Cordilleran till refers to any till deposited by glaciers that emerged from the Western Canadian Cordillera. In the Western Plains portion of the study area the relatively fine texture of the till, coupled with the more subdued relief, result in large areas of organic soil. Thirteen per cent of the 83 C/16 (Blackstone River) mapsheet area is mapped as organic terrain, reflecting poor drainage. Well drained sites on the plains and along major valleys support Grey Luvisols but Brunisols are generally more common, especially in areas of greater relief within the Foothills. The steeper slopes of the Foothills and the mountains are subject to rapid rates of denudation and therefore soils of the Regosolic Order prevail.

The high carbonate content of the glacial deposits frequently results in *Cca* soil horizons, particularly in coarser-textured materials. Reimchen and Bayrock (1977) used the depth of carbonate leaching as an indicator of relative age of deposits along the Foothills. A discussion of this approach is presented in section 4.2.6. Field descriptions of the soils were obtained at all sample sites. The depths of the solum and each major soil horizon were recorded. Notes were made of the parent material, the observable presence of ash or loess, the color and structure of the major horizons and the surface vegetation. The profile descriptions were based on The System of Soil Classification for Canada (Canada Dept. of Agriculture, 1974) except that the soil colors were defined according to the Revised Standard Color Charts (Oyama and Takehara, 1967). Textures were determined by hand rubbing the samples. The carbonate content of the soils was assessed visually and by observing the reaction with dilute hydrochloric acid. The presence of iron stains, mottling and weathering rinds on clasts was noted.

1.3.6 LAKE AND BOG CORE ANALYSIS:

A piston corer similar to the Livingston model (Cushing and Wright, 1965) was fabricated. The core barrel consisted of a 5.08 cm (2 in) diameter stainless steel pipe, one metre in length. Cores were obtained from three bogs and two lakes within the study area. The bogs were cored in autumn when the water table was at its lowest and the lakes were cored during the winter so that the coring could be carried out from the ice surface. All of the cores were measured, described and sampled before freezing to reduce the amount of sediment disturbance. The cores were wrapped in thin plastic and aluminum foil as soon as they were extracted. The first cores obtained were returned to the laboratory on corrugated fiberglass racks but it was found that they could be better protected and insulated if they were individually wrapped in single-sided corrugated cardboard and stacked in sturdy cardboard boxes. In the laboratory the sediments were sampled at 5 cm intervals along the length of the core. These samples were placed in clean plastic containers and frozen for pollen analysis (see Section 4.3).

The lakes that were cored were selected because they are sustained by local catchment areas rather than by streams. In addition, they are sufficiently deep that the bottom sediments are unlikely to have been disturbed by large animals, the sedimentation rates are probably slow and the lakes were almost certainly persistent through the Altithermal interval of Holocene time. The Muskiki Lake cores were extracted from beneath 7.5 m of ice and water and the depth of Fairfax Lake was 5.8 m at the location where the coring was carried out. These lakes represent the eastern (Fairfax) and western (Muskiki) sides of the Foothills Belt. The primary purpose of obtaining the lake and bog cores was to acquire basal dates from the lowest sediments. Radiocarbon dates from the samples were determined by the Radiocarbon Dating Laboratory of the Saskatchewan Research Council. The samples sent for analysis consisted of a 10 cm slug of the core, subdivided into contiguous upper and lower 5 cm slugs. Instructions were given to the laboratory staff to use only the lower slug if sufficient organic carbon could be obtained and to add the upper slug if the carbon content was too small for an accurate date determination. In all cases only the lower slug was required.

The high incidence of calcareous bedrock within the study area (see Fig. 1.2) poses a potential threat of contamination by 'dead' carbon (Clayton and Moran, 1982).

Attempts were made to minimize this by testing the adjacent portions of the core with hydrochloric acid to check for the presence of carbonate material. Sections of the cores which registered relatively high carbonate contents were avoided for dating purposes. The laboratory staff were also advised of the possible contamination problem so that they could pre-treat the samples. The fact that the dates obtained are all somewhat younger, rather than older, than expected suggests that this source of contamination was not a major factor.

A steel probe was used at the coring sites to verify that the cores included a cross-section of all sediment within each basin. For the Muskiki Lake site there is some question as to whether or not the sediments extend deeper than the bottom of the core because a definite contact with bedrock could not be determined by probing. Fairfax Lake had a more obvious contact between the consolidated and unconsolidated sediments. Similarly, the bottom of the organic bog deposits was relatively easy to verify.

1.3.7 TEPHRA ANALYSIS:

The lake core sediments contained identifiable tephra layers which were useful marker horizons. Two ash layers from the Fairfax Lake core were subjected to analysis using the A.R.L. "EMX" electron microprobe of the Department of Geology, University of Alberta. This technique compares the chemical composition of the volcanic glass shards to standard mineral samples with known mineral composition. The mineralogy of the pyroclasts can thereby be determined and the data obtained may be compared to values obtained from tephras of known origin.

The technique used closely follows that described by Smith and Westgate (1969). The tephra was wet-sieved through a 230- mesh (0.063 mm) sieve. The residue was discarded. Other researchers (Smith and Westgate, 1969; Westgate *et al.*, 1970) have demonstrated that the pyroclasts of Mazama, Glacier Peak, Bridge River and St. Helens 'Y' tephras are all of sub-sieve size.

Clear glass shards are required for the analysis. These were separated using liquids with a specific gravity of greater than 2.8 (methylene iodide and acetylene tetrabromide). A light fraction consisting of pumice with bubble inclusions, and lighter minerals, was separated and discarded. A second separation of material heavier than the

shards was removed by floating the shards. The specific gravity of the liquid was empirically adjusted by dilution with acetone. This technique was repeated until a relatively pure sample of shards was obtained. The shards were then mounted in epoxy resin and the sample was cut and polished using standard thin-section preparation techniques. A thin coat of carbon was deposited on the surface of the sample, after polishing, to render the surface conductive.

The samples were mounted in the electron microprobe machine and a beam of electrons was directed onto the surface of the shards for a period of 5 seconds. Ten shards were analyzed from each tephra sample. An analog count of the emitted radiation (photons) was maintained during the bombardment. This provided the basic data for the comparisons with samples of known composition which enables the mineral composition to be determined. Smith and Westgate (1969) demonstrated that as few as three elements (Ca, K, and Fe, or Ca, K, and Na) will provide sufficient evidence to discriminate between the tephras found in western Canada, but nine elements (Na, Mg, Al, Si, Cl, K, Ca, Ti, Fe) were used in this study, partially to familiarize the author with the techniques involved in their identification.

2. LITERATURE REVIEW

2.1 STUDY AREA

The earliest work which deals specifically with the surficial geology of the study area is an Alberta Research Council Report (No. 9) produced by Allan and Rutherford in 1924. The focus of that report pertains to coal resources but a short discussion of Quaternary deposits is included. Some of their observations were very astute. They noted, for example, that two ice sheets were responsible for the landform assemblages described. They considered that a Cordilleran Ice Sheet extended across the Foothills as an expanded piedmont system. This issued from the Brazeau, and other large valleys, and overtopped all of the Foothills in this area. A continental ("Keewatin Sheet") lobe interacted with the piedmont ice such that;

> "...the distribution of these (Cordilleran) deposits in the outer foothills and the western edge of the plains was influenced by the latter ('Keewatin') ice." (Allan and Rutherford, 1924,p. 38).

They also described "moraine ridges" (eskers) within the Brazeau Valley near the confluence of the North Brazeau (Cardinal) River and the Brazeau River. Limestone erratics 185 m up the valley walls through Cardinal Ridge were described, as were terraces along the Brazeau River valley east of the Foothills.

Reimchen and Bayrock (1977) studied the surficial geology and erosion potential of the Foothills of Alberta north of 52 degrees North Latitude. The study combined an airphoto interpretation with an extensive field investigation, by road and helicopter traverses, of the entire area. Samples were collected and analyzed by the Research Council of Alberta. The report includes descriptions of the surficial deposits and landforms plus a map of the surficial deposits (scale 1:250,000).

The report contains some inconsistencies. For example, the relative age of till units is defined according to the depth of leaching of calcium carbonate. Thus, *Young Till* is defined as having been leached 0'–2' (0–60cm), *Intermediate Till* is leached 2'–5' (60–150cm) and *Old Till* is leached to deeper than 5' (150cm). After defining the tills in this manner Reimchen and Bayrock (1977) outline general till characteristics. For example they state that the *Old Till* is "...thin, being on the average 5 feet in thickness" (Reimchen

and Bayrock, 1977, p. 9). However, if the *Old Till* averages 5 feet (150cm) in thickness, all of the *Old Till* thinner than that average depth strictly should not be classified as *Old Till* because the defined depth of leaching exceeds the thickness of the deposit.

Three Cordilleran advances and two Continental advances were hypothesized by Reimchen and Bayrock (1977, p.52) for the Foothills region between the Brazeau River and the Red Deer River. Distinctions between the deposits produced by these events were based on:

(1) the degree of leaching

(2) the degree of dissection

(3) geomorphic associations of the deposits.

For the area, between the Athabasca River and Robb, the geologic evidence was interpreted as reflecting a fourth ("ancient") Cordilleran advance. They suggested that the Late Wisconsin Cordilleran advance probably deposited sediments of two different ages but these were not differentiated in the report.

According to Reimchen and Bayrock (1977), the oldest Cordilleran deposits were found at high elevations (above 4200 ft, 1377 m) on the eastern slopes of the Brazeau and Ram Ranges, and the youngest deposits were restricted to the major river valleys and extended eastward "...beyond the edge of the Foothills" (Reimchen and Bayrock, 1977, p. 42). They added that "The youngest continental glaciation has glacial sediments interfingering with glacial sediments of the youngest Cordilleran glaciation in the Rocky Mountain Foothills" (Reimchen and Bayrock, 1977, p. 42).

2.2 STUDIES OF ADJACENT AREAS

The Athabasca River valley is a major drainage system north of the study area. Roed (1968, 1975) mapped and interpreted the glacial deposits along this valley from the mountain front to the eastern side of the area depicted by the Edson–Hinton map sheet (NTS 83F). The Foothills were excluded from his study. Informal rock stratigraphic units were defined for the surficial deposits. Roed (1968, 1975) interpreted, described and named seven tills, three gravel units and one inter–till outwash unit. In addition he mapped and described several glaciofluvial and glaciolacustrine deposits within the area. He interpreted all of these deposits as indicative of a glacial record which spans the

geologic period from Late Tertiary time to the Early, Middle and Late stages of the Pinedale Glaciation (Classical Wisconsin). The deposits were not dated radiometrically or by relative age-dating techniques other than stratigraphic position. The assigned ages for these units were obtained by correlating their stratigraphic position with other dated units in the province and in the Rocky Mountains of Wyoming. Except for indirect evidence of an *Early Cordilleran* glaciation no explanation was provided for the large hiatus between the Tertiary/early Quaternary gravel deposits and the Wisconsin-equivalent glacial units. Figure 2.1 and Table 2.1 depict the glacial events and the relationships between Cordilleran and Laurentide ice suggested by Roed (1968, 1975). On the basis of stratigraphic relationships and relative magnitude of events he correlated his work with that of St.-Onge (1972), Boydell *et al., (1974)* and Rutter (1977), as illustrated by Table 2.1.

To the east and south of the area covered by the present study lie the drainage systems of the Nordegg, North Saskatchewan and Ram Rivers. This area (NTS 83B mapsheet) was investigated by Boydell, Bayrock and Reimchen (1974) and described by Boydell (1972, 1978). Boydell (1972, 1978), like Roed (1968, 1975), placed a major emphasis on stratigraphic and lithologic evidence, in the absence of radiometric dates for the glacial events which he described. Nevertheless, a chronology based on relative dating techniques, primarily the superimposition of deposits, was outlined. The areas above elevations of 1660 m (mainly the Ram and Brazeau Ranges) were thought to have escaped glaciation during Late Wisconsin time, although quartzite and conglomerate erratics indicated that at some time they had been covered by active ice (Boydell, 1972, p. 20). Figure 2.2 illustrates a simplified model of Boydell's (1978) interpretation involving four glacial events, three of which were Cordilleran. On the basis of stratigraphy and relative position, Boydell (1978) related his work to that of Roed (1968), McPherson (1970) and Boydell (1970; 1972) as illustrated by Table 2.2. Boydell (1972) interpreted the Athabasca till (which he associated with the Foothills Erratics Train) as the product of a Laurentide advance. According to Roed (1975), he adopted the position that the Athabasca deposits were of Cordilleran origin and were deposited by ice flowing southeastward along the edge of the Foothills. He later noted (Boydell, 1978) that the lithologic evidence suggested a mixed Cordilleran and Laurentide origin.

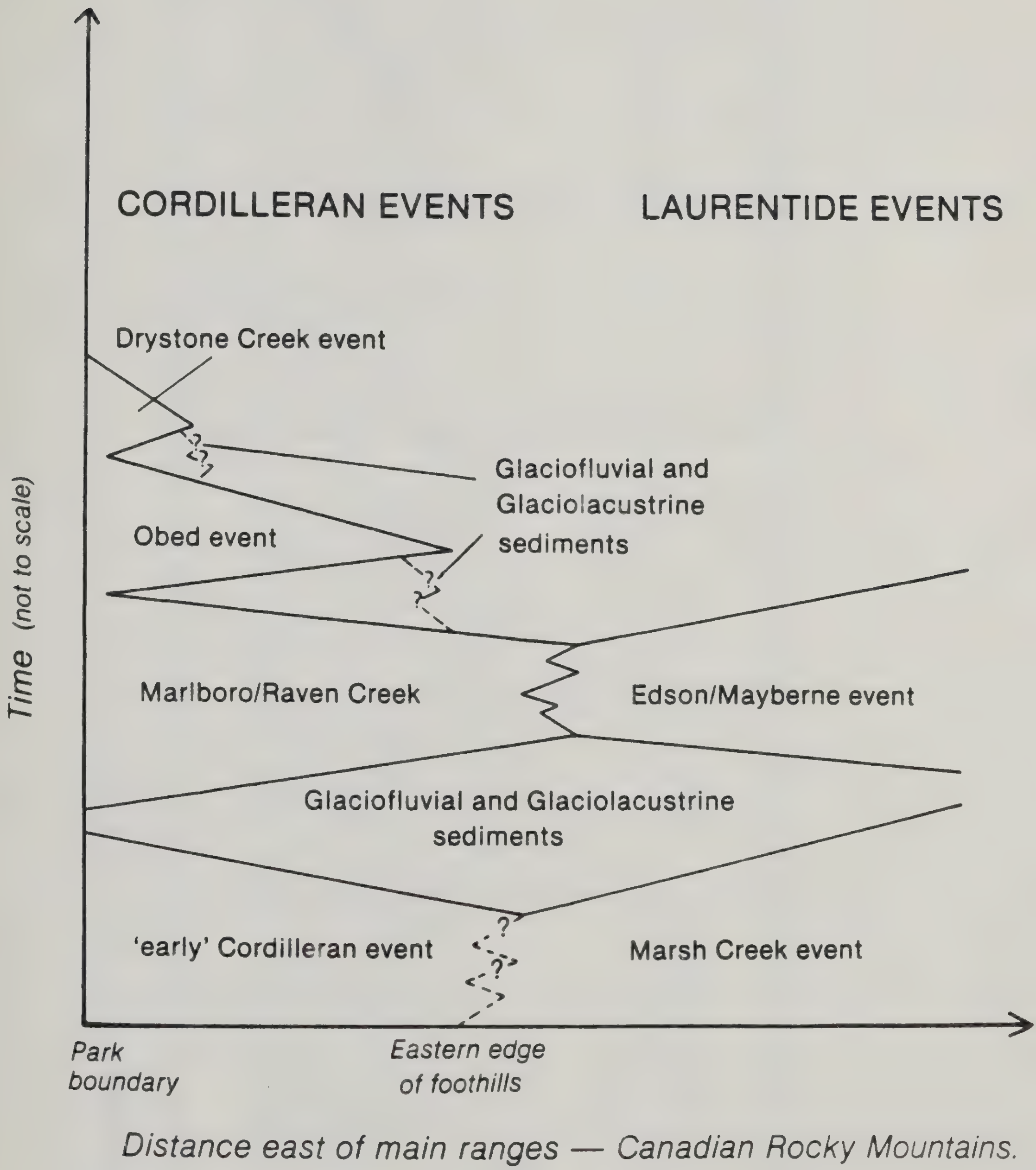


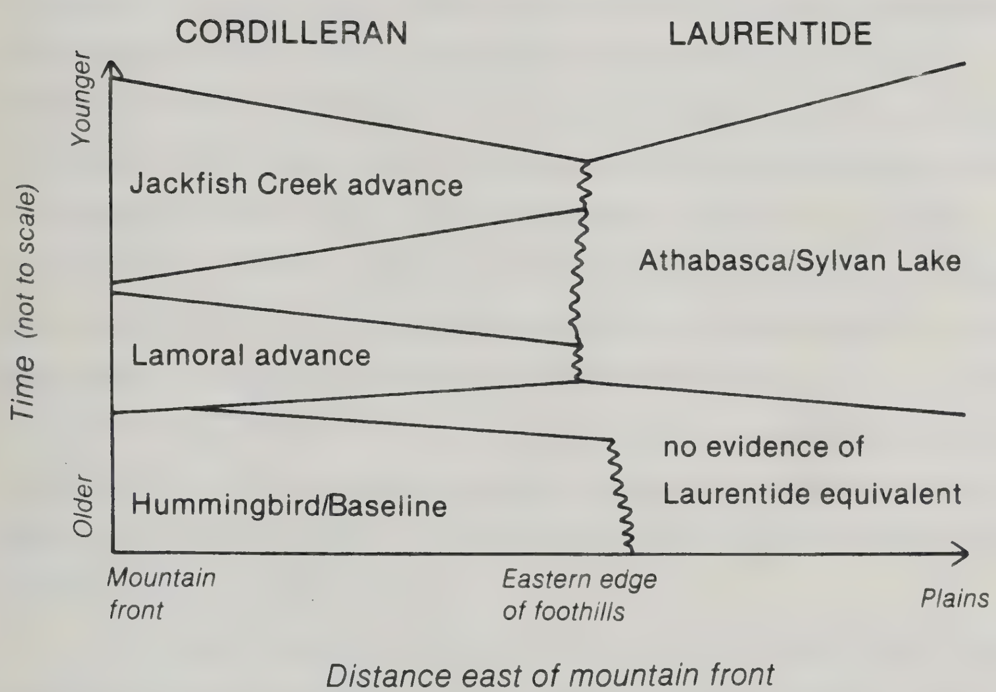
Fig. 2.1 Glacial Events, Hinton — Edson Area Source: M.A. Roed, 1975

Table 2.1 Intercorrelation Chart after Roed, (1975).

	Cordilleran origin				Laurentide origin		
Author	Rutter (1972)	Boydell (1978)	Roed (1975)	Rutter (1977)	Boydell (1978)	Roed (1975)	St.-Onge (1972a,b)
Area	Banff	Rocky Mtn. House	Edson-Hinton	Williston Lake	Rocky Mtn. House	Edson-Hinton	North-central Alberta
Name of till	Eisenhower Junction		Drystone Creek	Deserter's Canyon			
	Canmore		Obed	Late Portage Mtn.			
	Bow Valley	Lamoral-Jackfish and Athabasca	Marlboro-Raven Creek	Early Portage Mtn.	Sylvan Lake	Edson-Mayberne	Upper Till
						Marsh Creek	Lower Till

Table 2.2 Intercorrelation Chart after Boydell, (1978).

		Edson-Hinton area Roed (1968)		Sundre-Red Deer valley area Boydell (1970)		Upper North Saskatchewan River valley McPherson (1970)	Rocky Mountain House area Boydell (1972)	
Age		Rocky Mountain	Laurentide	Rocky Mountain	Laurentide	Rocky Mountain	Rocky Mountain	Laurentide
LATE WISC.		Drystone Creek	not present	not present	not present	not present	not present	not present
		Obed	not present	Elkton Creek	Sundre	Main	Jackfish Creek	Athabasca-Sylvan Lake
		Marlboro	Edson	not present	?	Big Horn	Lamoral	
EARLY WISC.		Valley or piedmont (?)	Marsh Creek	?	?	?	Baseline	?
		Early Cordilleran (?)	?	?	?	?	Hummingbird	?



Source: A.N. Boydell, (1978)

Fig. 2.2 Glacial Events, Rocky Mountain House Area

Immediately west of Boydell's (1972) study area, and the next major drainage system south of the present study area, is the section of the North Saskatchewan River which was mapped and interpreted by McPherson (1970). McPherson's (1970) study focused on the landforms and surficial deposits along the North Saskatchewan River valley within the Main and Front Ranges of the Rocky Mountains. He suggested that Bow Pass acted as an ice divide during the last major glaciation as there is no lithologic evidence of a flow of ice from west of the Continental Divide. One radiocarbon date (Westgate and Dreimanis, 1967) was cited and this was interpreted as a minimum date for deglaciation of the upper portion of the valley at 9330 +/− 170 (GSC - 332) years B.P. (McPherson, 1970, p. 18). The stratigraphy within the North Saskatchewan River valley consisted of two units of till separated by a thick (more than 75 m) unit of sand and gravel. This stratigraphic sequence was interpreted as evidence of two ice advances, the *Big Horn* and the *Main* advances. On the basis of its stratigraphic position, and the single radiocarbon date, the *Main* advance was thought to represent the Late Wisconsin advance of the Mid−Continent and the Big Horn advance could have been an earlier Wisconsin or a pre−Wisconsin event.

Transecting the eastern edge of the present study area is a train of quartzite and pebbly quartzite erratics which has been mapped from the Athabasca Valley to south of the 49th parallel. The erratics were noted and described by a number of early geologists, beginning with Hector of the Palliser Expedition of 1863 (Spry, 1963) but Stalker (1956) was the first to do a systematic study of the erratics train. It was his report which drew attention to the fact that the erratics form a continuous train along the eastern margin of the Foothills from 53 degrees 30 minutes North Latitude to the Canada − United States Border. He interpreted the source of the erratics as being either the Precambrian Shield or Lower Cambrian units within the Rocky Mountains. Stalker (1956) noted that Laurentide ice had at some time extended further west than the erratics train because crystalline Shield erratics were found west of the train. He suggested that the location of the train was influenced more by continental than mountain ice. Because of the lack of road access Stalker (1956) did not map erratics within the area covered by the present study. Mountjoy (1958) extended Stalker's (1956) work by determining that the Jasper area was the most probable source for the quartzitic erratics.

Roed *et al.*, (1967) conducted a study of an erratics train which they called the Athabasca Valley Erratics Train. They demonstrated that this train of erratics is comprised of schists, rather than quartzite, and they found that the source area for the erratics was most probably the Monashee Mountains and/or the Premier Range of British Columbia. The significance of this is that it illustrates that ice flowed across the Continental Divide at some time during a relatively recent glacial event. Both Stalker, 1956, p. 12, and Roed, *et al.*, 1967, p. 627, noted that most of the ice was carried out of the mountains via the Athabasca Valley and erratics have not been traced back to the Brazeau, North Saskatchewan or Bow River valleys. However, these authors note that this is a tentative conclusion and future research may indicate that the origin(s) of these erratics requires a relatively complex explanation. This issue is discussed in more detail in Section 6.4 of this study.

2.3 PLEISTOCENE AND HOLOCENE STUDIES FROM THE MAIN RANGES IMMEDIATELY WEST OF THE STUDY AREA

Luckman *et. al.*, (1978) and Luckman and Osborn (1979) studied the geomorphological and chronostratigraphic evidence for late-Pleistocene and Holocene glacial advances in the Main ranges of the Rocky Mountains. They employed relative (relative position, lichenometry), stratigraphic (tephrochronology) and absolute (C 14, dendrochronology) dating techniques to assist their interpretation of the late-glacial and post-glacial chronology.

In essence, they support earlier conclusions by McPherson (1970) and Rutter (1972) that late Wisconsin deglaciation in the area was dominantly the result of stagnation and downwasting (Luckman and Osborn, 1979, p. 57). Radiocarbon evidence from two sites near the Continental Divide indicate that these mountain valleys were essentially ice free by about 10,000 B.P. (9660 +/- 280 years B.P. BGS-465 and 9600 +/- 300 years B.P. BGS-490) (Luckman and Kearney, 1979). They conclude that the last Cordilleran ice to occupy the mountain valleys in this area was part of a Cordilleran glacier which traversed the Rocky Mountains from a source region in the Interior of British Columbia. They feel that it was probably this event which produced the Athabasca Valley Erratic train described by Roed *et al.*, (1967) (Luckman and Osborn, 1979, p. 57).

Luckman and Osborn (1979) describe evidence for two minor glacial advances which were subsequent to the last major (late-Wisconsin?) advance. These they identify as the Crowfoot and Cavell advances. Both of these advances were confined to high alpine environments within a kilometer of contemporary valley glaciers. The Crowfoot deposits are overlain by tephra which was identified as a product of the Mazama eruption. This allows the Crowfoot deposits to be differentiated from the more recent Cavell advances.

From the perspective of the present study, a major contribution of Luckman *et. al.*, and Luckman and Osborn (1979) is the discussion they provide concerning the Chateau Lake Louise event (Harris and Howell (1977) cited in Luckman and Osborn (1979, p. 71), the Pinedale IV events of Reeves and Dormaar (1972), and the Deserters Canyon Advance (Rutter 1976, 1977). The dating control on these events is at least suspect according to Luckman and Osborn (1979) and the weaknesses are clearly identified. They suggest that the morphological evidence employed by Harris and Howell (1977) is equivocal and the age assigned to the surface is only a minimum age and therefore only indicates that the surface is older than the 8800 yrs. B.P. age which has been assigned to it. The criticism of the Pinedale IV and subsequent events (Reeves and Dormaar, 1972) is that Mazama tephra has been identified on cirque moraines and in cirque lake cores within Waterton Park (Luckman and Osborn, 1979, p. 72). This evidence, together with the absence of morphological evidence of terminal ice positions in the areas identified by Reeves and Dormaar (1972), casts doubt on the Pinedale IV and subsequent glacial events. The Deserters Canyon Advance (Rutter, 1976, 1977) is challenged in two ways. From a regional perspective it is difficult to explain why a glacier would extend for more than 150 km down the Peace River Valley at a time when the valleys adjacent to the contemporary icefields were ice-free. Luckman and Osborn's (1979) second concern is the absence of a demonstrated connection between the terrace sediments, from which the material used to date the event was obtained, and the Deserters Canyon till sheet which provides a major line of evidence for the event.

2.4 PLEISTOCENE GEOLOGY STUDIES NORTH OF STUDY AREA

The portion of northeastern British Columbia which lies within the Interior Plains physiographic region (Bostock, 1948), and is drained by the Peace River system, has been extensively studied by a number of researchers. The published results of these studies may be traced back to Dawson (1888a,b; 1890a,b;) and McConnell (1896). Rutter (1977, p. 1-3) provides a review of this literature. A problem which many of the researchers address (Dawson, 1888; Kerr, 1934; Armstrong and Tipper, 1948, and Tipper, 1971a,b) pertains to determining the nature of the ice mass which formed within the Western Canadian Cordillera. In particular, those authors were trying to determine whether or not a Cordilleran Ice Sheet was established in north-central British Columbia during the last major glaciation. Some aspects of this problem remain unsolved (Kvill, 1976). There is general agreement, however, that at least two Cordilleran advances occurred within the area.

Mathews (1954, 1962, 1963, 1980) mapped and described the Quaternary stratigraphy and geomorphology of the Fort St. John area which lies approximately 80 km east of the Rocky Mountains. The stratigraphic evidence provides a strong argument for two major glacial events because the glacial deposits are separated by nonglacial sediments. According to Mathews (1963) a continental ice advance invaded the area and obstructed the normal drainage at some time prior to the last advance. The maximum extent of this early advance is not known.

The most recent advance to affect the area is well marked by geomorphic and stratigraphic evidence. Mathews (1963, p. 14) stated that:

> "During at least its *(Laurentide ice)* latest period of activity this ice moved into the Fort St. John area from the northeast and flowed in a broad sweeping curve southwest, south, then southeast ... as if deflected by the front of the Rocky Mountains or by ice tongues issuing from it."

Mathews (1963, 1980) argued that the Cordilleran ice of the last advance appeared to have overridden areas from which the Laurentide ice had already withdrawn. The maximum extent of this Cordilleran event resulted in that ice extending about 56 km beyond the mountain front. Glaciofluvial and glaciolacustrine sediments, as well as meltwater channels, are documented at high elevations indicating that extensive ponding

occurred between the two major ice masses during the early stages of deglaciation.

In a later paper Mathews (1980) provides additional data on the Fort St John area. He presents a series of maps depicting the phases of deglaciation from an initial stage of coalescent Laurentide and Cordilleran ice to the final withdrawal of Laurentide ice from the area. In this study Mathews (1980) divides the eastward flowing ice into Cordilleran ice and Mountain ice on the basis of lithologic and geomorphic evidence. For the purposes of this study the distinction is not particularly important and Mountain ice can be considered a sub-set of Cordilleran ice. Mathews (1980) presents fourteen radiocarbon dates which he cites as evidence that the last major glacial event in the area occurred between 25,000 and 10,000 years B.P.

For the Dawson Creek area Reimchen and Rutter (1972) described evidence of three Cordilleran and two Laurentide advances. The stratigraphic evidence was interpreted as indicating an old Cordilleran advance, the deposits of which have been weathered to depths of 10 to 14 feet (3 to 4.25 m) (Reimchen and Rutter, 1972, p. 92). This advance may have merged with Laurentide ice and the Laurentide margin was at least sufficiently close to obstruct local drainage. One, or possibly two, subsequent Cordilleran events were postulated by the authors and associated with a 15 mile (24 km) wide zone of mixed Cordilleran and Laurentide deposits. Drumlins formed by Cordilleran ice truncate those of continental origin. This led the authors to postulate that the related Laurentide advance predated the maximum Cordilleran advance (Reimchen and Rutter, 1972, p. 92). Roed (1975) and Mathews (1980) also use abrupt changes in ice-flow indicators as evidence of temporally separate events. Caution is advised in such interpretation because it is at least theoretically possible for an abrupt change in ice flow from coalescent glaciers to simply reflect a change in the dynamic balance between the two glaciers. For example, if accumulation amounts in Cordilleran valley glaciers remained stable, the distal margin of this glacier was in contact with Laurentide ice, and if the supply of ice to the Laurentide margin was to diminish, the level of the Cordilleran ice would drop and the direction of ice flow would reflect the increased influence of topography.

Gastropod shells from glaciolacustrine sediments deposited during deglaciation have furnished a minimum date for deglaciation of 16,300 +/- 180 years B.P. (GSC-1548). However this date has since been declared erroneous and recalculated as

9960 +/− 170 years B.P. by Rutter (1977).

Rutter (1977) continued the investigation of the Quaternary geology of this region by mapping and describing the Williston Lake area. This study represents a significant contribution to the problem of establishing a regional glacial chronology for the eastern slopes of the Rocky Mountains because eleven radiocarbon dates were presented. These provided a more accurate basis for his chronology. Rutter (1977) found that the oldest sediments within the area were gravel and sand deposits which he interpreted as interglacial fluvial sediments. Overlying these were glacial and glaciofluvial sediments of the *Early* glacial advance. Sediments from this event were found to be relatively extensive and separated by glaciofluvial and glaciolacustrine sediments from deposits of the *Early Portage Mountain* glacial event. These were traced east to beyond the mountain front. The surficial deposits east of the mountains indicated that ponding occurred between Cordilleran and Laurentide ice during this event. Deposits of a later glacial event (*Late Portage Mountain*) were traced through the Peace River valley to the eastern edge of the mountains where they terminated at a large kame moraine. Evidence from the mountain valleys indicated that a minor glacial event (*Deserters Canyon*) occurred after deglaciation of the *Late Portage Mountain* event.

The radiometric dates were obtained from organic material associated with some of the stratigraphic units. Wood and peat from the oxidized gravel and sand deposits beneath the *Early* till furnished dates older than could be measured by conventional radiocarbon dating methods. Plant material, collected from glaciolacustrine sediments thought to be stratigraphically linked to the *Early* till, produced a date of 25,940 +/− 380 years B.P.(GSC−537) for deglaciation of this event. This date indicated that the *Portage Mountain* events were almost certainly of Late Wisconsin age. A mammoth tusk excavated from the sediments of the *Late Portage Mountain* kame moraine furnished a radiocarbon date of 11,600 +/− 1,000 years B.P. (I−2244A). Mathews (1980) questioned this date calculation on the basis of the small amount of collagen that was used. He also pointed out that an earlier attempt to date the same sample furnished a date of 7670 +/− 170 years B.P. (I−2244) on the carbonate fraction. A more recent attempt, utilizing a new sample from the tusk, produced a date of 25,800 +/− 320 years B.P. (GSC−2859) but this date may have been contaminated by a preservative (Mathews, 1980, p. 19).

A date of 9,280 +/- 200 years B.P. (GSC-1497) from bone found in ice-contact sediments established a date for the late-retreat stage of the *Late Portage Mountain* advance. Dates of 7470 +/- 140 years B.P. (GSC-1069) and 7470 +/- 150 years B.P. (GSC-1161) from wood and peat found in terrace sediments provide minimum dates for deglaciation of the *Deserters Canyon* event.

Further north, along the eastern portion of the Rocky Mountain system, Ford (1976) compiled a glacial chronology based on landform and speleothem evidence. He concluded that during the Late Wisconsin advance Laurentide ice entered the eastern edge of the Cordillera and dammed the South Nahanni River to create Glacial Lake Tetcela. Similar evidence was found for an earlier, probably Illinoian, (Ford, 1976, p. 1442) advance which created a more extensive lake (Glacial Lake Nahanni). Ford (1976) found no evidence of cover by an ice sheet of regional extent or of large valley and transectional glaciers in the Funeral, Headless or Tlogotsho ranges and the western half of the Nahanni Plateau. All of these areas lie near the eastern edge of the Cordillera.

The pattern of Laurentide ice abutting against, or penetrating, the mountain front was also documented further north in the Mackenzie and Richardson Mountains by Hughes (1969). Evidence of Laurentide ice activity has been traced to elevations of about 1070 m on the eastern slopes of the Richardson Mountains and 1220 m on the northeastern slopes of the Mackenzie Mountains. Hughes (1969) relied heavily on geomorphic evidence in his work and cited numerous examples of moraines, meltwater channels and glaciofluvial features. In addition, he used the relative preservation of landform morphology to distinguish between surfaces of different ages.

Evidence was found of three Cordilleran and two Laurentide advances. However there was no evidence of coalescence of the two ice masses during any event. A number of radiometric dates were cited but these come from an extremely extensive area. Hughes (1969) felt that the dates suggest that deglaciation along the eastern slopes of the Richardson Mountains occurred by approximately 9970 +/- 180 years B.P. (GSC-147). Minimum dates for the intermediate-aged valley glaciation event were obtained from frozen bog sediments. Two dates (12,550 +/- 190 years B.P. (GSC-128) and 13,780 +/- 180 years B.P. (GSC-296) were cited but these samples were collected from areas in the western Yukon which are only correlated by relative position with the

landforms of the eastern slopes. One date from sediments beneath Laurentide till, near the terminal position of the last advance, was collected but the organic material proved to be beyond the maximum age of conventional radiometric techniques. It is therefore only useful to illustrate that this material is at least 50,000 years old.

2.5 PLEISTOCENE GEOLOGY STUDIES FROM SOUTHWESTERN ALBERTA

The superior road access, relatively light forest cover, climate and excellent stratigraphic sections combine to make southwestern Alberta a popular area for Quaternary studies. As a result, numerous relevant theses and publications have accumulated. For brevity only selected works will be discussed here. This does not imply that the omitted studies are less important but rather that they are incorporated in, or are consistent with, the published results of the reviewed studies.

Work in the Bow River valley by Rutter (1965, 1966, 1972) represents the first studies pertaining exclusively to glacial geology in that area. Stratigraphic evidence for three, and possibly four, advances was presented and the suggested chronology was correlated with that developed by Blackwelder (1915) and Richmond (1965) for areas of western United States. Using the revised nomenclature presented in Rutter (1972), the sequence of events described for the Bow River valley is outlined as follows:

(1) A till unit and extensive outwash unit were considered to indicate an early advance of Cordilleran ice, the maximum extent of which was not determined. Deposits of the subsequent *(Bow Valley)* advance overlie these sediments. The latter event was interpreted as the most extensive event for which good stratigraphic evidence is available. Breaks-of-slope, erratics and ice-contact deposits mark the distal limit of the retreat phases of this advance. Rutter (1966, p.61) felt that this advance probably extended well to the east of the Kananaskis River valley. This implies that the advance probably extended east of the mountain front as well. No end moraines were found and the tills of the Bow Valley advance were indistinguishable from the younger *(Canmore)* advance so the actual extent of the Bow Valley advance was interpreted from indirect evidence such as glaciofluvial and ice-contact deposits.

(2) The *Canmore* advance may represent a readvance of the Bow Valley glacier (Rutter, 1972, p.22). The stratigraphic evidence indicates that the Bow Valley ice retreated

upvalley as far as Banff, then readvanced an unspecified distance downvalley.

(3) The last advance, according to Rutter (1972), was the *Eisenhower Junction* advance which, in contrast to the earlier events, was defined on the basis of lateral and terminal moraines, as well as breaks-of-slope and ground moraine. As the name suggests, the maximum extent of this advance terminated near Eisenhower Junction on Highway #1. During this event ice extended downvalley to, but not into, the Front Ranges of the Rocky Mountains. No radiometric dates were presented for these events, except by association with the Saskatchewan River Crossing area where a date of 9330 +/- 170 years B.P. (GSC-332) (Westgate and Dreimanis, 1967) is used as an approximate date for deglaciation after the Eisenhower Junction event.

Smith and Harris (1977) also presented a glacial history for the Bow River valley comprised of three Cordilleran advances but some of Rutter's earlier interpretations were harshly criticized. Smith and Harris state that during the last major advance "...Mountain ice reached Calgary, as did Laurentide Ice" (Smith and Harris, 1977, p.12). Shells from glacial lake sediments near Cochrane (30 km east of the mountain front) have been radiometrically dated at 9650 +/- 160 years B.P. (I-5676). This evidence was used to support the argument that the Late Wisconsin ice was retreating up the Bow River Valley at approximately this time. Jackson *et al.*, (1982) have challenged this interpretation. They provide sedimentological, palynological and radiometric data which suggest that the Cochrane terraces were formed by paraglacial processes rather than glaciofluvial processes. Jackson *et al.*, (1982) conclude that the terraces were formed approximately 11,500 - 10,000 yrs. B.P. and by approximately 10,000 B.P. this portion of the Bow River valley was occupied by spruce and pine forests. Evidence was cited to indicate that remnant Laurentide ice remained in the Calgary area until about 9600 years B.P. (Smith and Harris, 1977, p.16). Smith and Harris (1977) also contended that the *Eisenhower Junction* end moraine identified by Rutter (1966, 1972) does not exist. Luckman and Osborn (1979) also question Rutter's (1966, 1972) evidence, particularly the dating control.

Harris and Howell (1977a) suggested that there is evidence for an early Holocene advance (Chateau Lake Louise) in the Bow Valley system. they assign a date of 8800 years B.P. to this event but Luckman *et. al.*, and Luckman and Osborn (1979) correctly

demonstrate that this date is only a minimum date and the event could have equally as plausibly been a late Pleistocene event. [1]

Jackson (1977) investigated the Quaternary deposits of the Kananaskis Valley which is tributary to the Bow River valley. In concordance with other studies in the area (Rutter, 1972; Smith and Harris, 1977) he found evidence of three major glacial events. The discrepancy between Jackson's (1977) study and Rutter's (1972) work is that Jackson identified the *Bow Valley* event as the first of three stades of the last *(Erratics Train)* glaciation. The *Canmore* and the *Eisenhower Junction* advances were interpreted as subsequent stades of this advance. The age of the *Erratics Train* event has been tentatively established as Early Wisconsin on the basis of a date obtained from wood found in what is believed to be the *Erratics Train* till (49,400 +/- 1000 years B.P. (GSC-2409).

The location of this discovery site is called Midnapore Bluff which is on the eastern side of the Bow River in the southernmost district of Calgary. At the Midnapore Bluff site Jackson identified and described a buried soil horizon which overlies the till of the *Foothills Erratics Train* glaciation (Jackson, 1977, 1979). The spruce log, from which the date mentioned above was obtained, was exhumed from a till unit which is thought to be the stratigraphic equivalent of this unit although apparently from another location several kilometers away (Jackson, 1976). No paleosol was mentioned at the site where the log was discovered.

To the southwest of Calgary, in the Foothills, Jackson (1979) discovered a deep bog in a meltwater channel which is thought to have formed by the glacier which deposited the Foothills Erratics Train (Stalker, 1956). This site (Chalmer's Bog) is located southwest of Turner Valley at an elevation of approximately 1370 m (Jackson, 1979).

[1] Subsequent to the original draft of this study, a fieldtrip of the International Association of Sedimentologists was conducted through the Rocky Mountains (Proudfoot *et al*, 1982). The excursion entered the mountains via the Athabasca River valley, proceeded southeast along the Banff-Jasper highway which generally follows the structural valley between the Front Ranges and the Main Ranges, and exited the mountains through the Bow River Valley. The Field Excursion Guidebook for this study contains a useful synthesis of most of the studies discussed above. Moreover, the interpretations expressed by the original authors, especially Rutter (1972) and Roed (1975), are presented without comments concerning the alternative interpretation and criticisms expressed by authors such as Shaw (1972), Harris and Howell (1977), Luckman and Osborn (1979) and Jackson *et al.*, (1982). The evidence upon which these interpretations are based is clearly described, however, which ensures that this study (Proudfoot *et al*, 1982) is a useful summary statement of a number of studies germane to glacio-stratigraphic studies in this area.

Two very important dates of 18,400 +/- 380 (GSC 2668) and 18,500 +/- 1,090 yrs. B.P. (GSC 2670) have been obtained from peat bands in a 12 m long core extracted with a Livingston piston corer. Jackson (1979) cites these dates and this site as evidence that the area was not overridden by the Late Wisconsin advance and thus the Canmore Advance (Rutter, 1965) probably marks the most distal extent of the last major Cordilleran advance in southwestern Alberta. The *Erratics Train* till described by Jackson (1977; 1979) and others has been interpreted as a mixed provenance till, containing Cordilleran and Laurentide lithologies, which suggests ice coalescence during this or a previous event. A number of other authors have documented a zone of mixed provenance deposits in the Calgary area and south along the eastern margin of the Foothills (Dawson, 1883; Coleman, 1909; Johnson and Wickenden, 1931; Stalker, 1956; 1961; 1973b; Morgan, 1966) and Proudfoot *et al.*, (1982) but none of these studies furnished radiometric dates to firmly establish their chronologies. The Bow River terrace study (Jackson *et al.*, 1982) corroborates the glacial chronology suggested by Jackson (1977).

Stalker has been investigating the glacial deposits of southern Alberta for more than two decades. Much of his Cordilleran work is synthesized in Stalker and Harrison (1977). That study, and Stalker (1977), draw a useful analogy between the relationship of Cordilleran and Laurentide ice masses and the two sides of a mechanical zipper. They suggested that during the maximum stage of each glacial event the ice masses first merged in the north and the zone of contact moved progressively southward. Subsequently, an ice-free area opened again from south to north. Stalker (1977) and Stalker and Harrison (1977) felt that the Cordilleran and Laurentide advances were slightly out of phase with one another so that the position of the closed portion of the zipper moved westward during the Laurentide maxima as the Cordilleran ice was retreating. The authors suggested that maximum coalescence occurred only during the earliest of four events. Each subsequent glaciation resulted in a smaller zone of ice contact, until the last advance when Cordilleran ice did not extend beyond the mountain front and the Laurentide ice failed to reach the Foothills in southern Alberta. The concept of nonsynchronous advances was noted in the earlier work of Alley (1972) and has been supported in a number of theses produced at the University of Calgary (see Harris and

Waters, 1977, for an extended discussion).

The recent work by Karlstrom (1981) in the Glacier and Waterton Parks area supports the conclusions reached by Stalker (1977) and Stalker and Harrison (1977). He uses pedological evidence to support his argument that Cordilleran and Laurentide ice has not merged in southwestern Alberta since at least Illinoian time. Karlstrom (1981) identified at least seven major glacial and interglacial events on the basis of the relative degree of soil profile development. The soils associated with post-Late Wisconsin deposits tend to be weakly developed whereas the post-Illinoian (pre-Wisconsin)soils are strongly developed, even in the mountains.

2.6 GLACIAL HISTORY WEST OF THE CONTINENTAL DIVIDE

Few studies have been carried out along the western slopes of the Continental Divide. Fox (1974) studied the glacial geomorphology of a valley in Yoho National Park but did not find datable material so his chronology of events is somewhat equivocal. He found evidence of four glacial events based on breaks-of-slope, lateral moraines and kame terraces. End moraines exist for only the last two of the events. No attempt was made to date the events except that the last advance was interpreted as a late Neo-glacial event and the others were all probably pre-Altithermal as the related deposits are overlain by a tephra which is assumed to be Mazama Ash.

Harrison (1976) discovered a 1 cm thick organic layer beneath Mazama (?) Ash in glacial lake deposits of the Elk River valley of British Columbia. Dates of 11,900 +/- 100 years B.P. (GSC-2142) and 12,200 +/- 160 (GSC-2275) years B.P. suggest that this area, which is only 24 km downvalley from an existing ice field, was deglaciated about 12,000 years ago. Harrison (1976) suggested that, because the Elk and Kananaskis Valleys shared a common ice source, this approximate date may also serve as a minimum date for deglaciation of the middle and lower Kananaskis Valley. Ferguson (1978) and Ferguson and Osborn (1981) present evidence from a 4.2 m long sediment core extracted from further up the Elk River valley from Harrison's (1976) study. They report a date of 13,430 +/- 450 years B.P. (GX-5599) from gastropod shells extracted from a clay zone which spanned the 3.6 to 4.1 m section of the "Weary Bog" core. Although the authors (Ferguson and Osborn, 1981) acknowledge some possibility of contamination of

this sample, extra precautions and sample preparations were taken to avoid the incorporation of old carbon in the sample. A most conservative interpretation would still have to recognize that this study augments the findings of Harrison (1976) discussed above. Pollen evidence from this sediment core reveals that a shrub tundra environment was to be found at the location 10 to 13 thousand years B.P. (Ferguson, 1978).

Clague *et al.*, (1980) presented a synthesis of 24 radiocarbon dates related to the inception and decay of the Late Wisconsin ice sheet in southern British Columbia. They suggested that much of the area, except for the high alpine zones, was not covered by Late Wisconsin ice until after about 19,000 to 20,000 years B.P. Then, in a very short period of time, the ice expanded to form a Cordilleran ice sheet which extended beyond the Canada – United States Border at about 15,000 years B.P. The rapid advance was mirrored by an equally rapid decay of the ice. In two millenia the Lower Mainland and the southern interior of British Columbia were again ice free (Clague, *et al.*, 1980, p. 325).

Smith (1979) studied the relative relief of the major mountain passes along the Continental Divide from southern Colorado to northwestern British Columbia. He found that on the western slopes of the passes the glaciofluvial and glaciolacustrine deposits indicated that ponding had occurred during the retreat of the last major glaciation. In order for ponding to occur on the western and not the eastern side of the divide, the ice source during the last glacial advance would have had to be centered west of the divide. Moreover, the fact that the sediments persist in these areas indicates that they have not been subsequently glacierized. In the case of the Yellowhead Pass, for example, the glacial event that produced the kettled topography and lacustrine sediments described by Smith (1979) must have been the Obed advance, or perhaps the Drystone Creek advance, of Roed (1968, 1975). If these deposits were formed by the earlier, and more extensive Marlboro advance (Roed, 1968, 1975), it is difficult to imagine why the ice of the Obed advance would flow east beyond the mountain front but not occupy the Yellowhead Pass area.

The radiometric dates presented by Harrison (1976) and Ferguson and Osborn (1981), noted earlier, indicate that ponding occurred along the western side of the Continental Divide in southeastern British Columbia about 12,000 years ago. This supports the contention that the glacial advance and subsequent retreat responsible for

the ponding took place during Late Wisconsin time.

2.7 GLACIAL CHRONOLOGIES FROM THE WESTERN MARGIN OF THE LAURENTIDE ICE

Although many of the studies from Alberta indicate that Cordilleran ice was not always in phase with the Laurentide advances, the fact that the Laurentide advances have been more extensively studied, and are stratigraphically simpler, makes it desirable to consider the evidence along the southwestern portion of the area covered by Laurentide ice.

St.-Onge (1972) traced the development of glacial lakes and, by inference, the frontal positions of the retreating Late Wisconsin Laurentide ice in north-central Alberta. The use of glacial lake sediments as a stratigraphic medium through which the two chronologies may be compared seems entirely reasonable, particularly when many of the Cordilleran studies discussed earlier cite evidence of extensive glaciolacustrine sediments along the distal margins of one or more end positions of Cordilleran ice (Mathews, 1954, 1962, 1963, 1980; Boydell, 1970, 1972, 1978; Rutter, 1977). St.-Onge (1972, p.8) suggested that Laurentide ice had retreated well east of the Foothills (Iosegun Lake area) by at least 13,520 +/- 230 years B.P. (GSC-694) and by 11,400 +/- 190 years B.P. (GSC-1049) much of central Alberta (at least as far east as Lofty Lake) was exposed. Lichti-Federovich (1970,1972) obtained radiometric dates on gyttja from a 5.5 m long core of sediment from Lofty Lake which substantiates these observations.

Christiansen (1979) provided a series of maps which depict the interpreted ice-frontal positions in Saskatchewan and easternmost Alberta during the retreat of Classical Wisconsin ice. Based on a synthesis of about two decades of fieldwork, and eleven significant radiocarbon dates, he concluded that the Classical Wisconsin advance was at its maximum extent sometime prior to 17,000 years B.P. At that time all of southeastern Alberta was thought to have been ice covered except for the higher elevations of the Cypress Hills, and possible nunataks along other higher areas of southwestern Alberta (e.g. Porcupine Hills). As St.-Onge (1980) suggested, Christiansen's (1979) chronology correlates very well with that of St.-Onge (1972).

Stalker (1977, Fig. 1, p. 2615) also synthesized a great deal of personal research in a map depicting his interpretation of the maximum extent of Laurentide ice in southern Alberta. The map indicates that during the Late Wisconsin advance Laurentide ice did not extend south of the Cypress Hills and did not cover a broad band of southern Alberta adjacent to the International Border. This conflicts with the interpretation of Christiansen (1979) mentioned above. Stalker (1977) also concluded that Cordilleran ice came in contact with Laurentide ice, as shown by evidence of a deflection in the Laurentide ice toward the southeast (Stalker's Innisfail Lobe). It is important to note that a western boundary was drawn for the Cordilleran ice which implies that south of the Athabasca River valley the Foothills were not ice covered. That hypothesis is discussed in Chapter 6 of the present study. The radiometric dates cited in Stalker's (1977) report are all from the Medicine Hat area and are only stratigraphically related to the ice-marginal positions of the Laurentide ice. Stalker's (1977) main line of evidence for the maximum extension of "Classical" Wisconsin ice is landform association. Most of the proposed ice margin identified on his map (Fig. 1, Stalker, 1977) is based on morainal evidence and observed differences in the physical appearance and drainage systems on either side of his boundaries. Stalker (1977) suggests that the area of least confidence is the western margin north of Innisfail. This is the area which is most critical to the present study.

Stalker and Harrison (1977, p. 888) note that the presence of deeply weathered (20 feet, 6 m) tills and extensively weathered boulders (rinds 1/2 - 2 cm thick) provide an important component of their argument for pre-Late-Wisconsin Cordilleran surfaces. This evidence, together with an extensive stratigraphic study of southern and southwestern Alberta, provides the basis for the interpretation that much of southwestern Alberta has been exposed since Illinoian time.

More recently, Clayton and Moran (1982) reviewed the late Wisconsin glacial chronology for the Laurentide advance in the central part of North America and Mickelson *et al.*, (in Wright, 1983) reviewed the evidence for the United States component of this sequence. In both studies it is apparent that considerable controversy still exists. The ambiguity which perpetuates this discussion centers around the paucity of stratigraphic or chronologic links between the ice marginal positions in southwestern Alberta and northwestern Montana (Cordilleran events) and those of southeastern Alberta

and northeastern Montana (Laurentide events). Reeves (1973) Stalker (1977) and Stalker and Harrison (1977) feel that late Wisconsin Laurentide ice did not extend into these areas, whereas the studies cited above (Clayton and Moran, 1982; Mickelson *et al.*, in Wright, 1983) feel that there is sufficient evidence to extend the Late Wisconsin Laurentide limit well into northern Montana. They provide evidence from Richmond (unpublished map), Madole (1980, cited in Wright, 1983 p. 9) and others, which documents early and middle Pinedale (approximately 20,000 yrs. B.P) alpine events, with the Late Wisconsin Laurentide expansion, on the basis of obstructed drainage systems in southern Alberta and northern Montana.

2.8 OTHER RELEVANT LITERATURE

A subject as broad as the glacial history of a regional ideally requires a multidisciplinary approach. Soils data, paleobotanical studies, archaeological evidence and paleoclimatic studies are all useful lines of evidence from which a model of glacial history may be constructed. Unfortunately, for the late Pleistocene environments of the general study area there is little information from other disciplines.

Packer and Vitt (1974) generated considerable excitement with their study of the Mountain Park area which is about 18 km northwest of the western boundary of the area covered in the present study. They identifed twelve disjunct plant species and interpreted these as evidence that the Mountain Park area was a plant refugium during the last glaciation. Disjunct species of crustaceans have been identified in the general area of Mountain Park (Clifford and Bergstrom, 1976; Daborn, 1976) and Pike (1978) identified three species of butterflies which are usually found only in Alaska, the Yukon or in the southern Rocky Mountains. It should be noted, however, that all of the disjunct species are not Arctic species and the existence of a refugium has not been firmly established as the only logical explanation for these disjunct species.

Peters (in press) mapped the soils of the Brazeau (83C/14) mapsheet area. He differentiated between parent material comprised of *Mountain* till and *Laurentide* till on the basis of Shield-derived clasts in the *Laurentide* till. These would approximately correspond to the Athabasca and Sylvan Lake tills identified as being of Late Wisconsin age by Boydell (1978). The contact between the tills of the two provenances is partially

masked by glaciolacustrine, aeolian and organic deposits but it generally follows an irregular, northwest-southeast trending, line which transects the Brazeau River near the damsite of the Brazeau Reservoir. Aeolian sand and lacustrine deposits are common throughout the area but are particularly prevalent along the boundary between the two till units. Peters (pers. comm.) suggested that most of the organic terrain overlies lacustrine sediments. This implies that lacustrine deposits are under-represented on the map. Peters (pers. comm.) did not observe soil profile differences which could be interpreted as evidence of older (pre-Late-Wisconsin) surfaces. His limited reconnaissance in the area further west (encompassed by the present study) failed to produce evidence of deeper or better developed soils in any of the locations which he observed. Nor was he able to identify more extensively developed soils in the upland areas, as compared to the lower elevations, as would be expected if the uplands were older surfaces as previously suggested by Reimchen and Bayrock (1977). Instead Peters (in press) found the better developed soils along the bottoms of the major valleys. Peter's (in press) evidence does not provide any insight on the age of these surfaces but it does support the concept that soils differences, as used by Birkeland (1974) and others, cannot provide the evidence needed to argue that these surfaces are of different ages.

No archaeological work has been published for the study area but the importance of the glacial history of the area to the anthropological question concerning the migration of early man into North America is paramount. The anthropological literature on the existence and significance of an ice-free corridor is too voluminous to review here but much of it is based on the hypotheses and evidence presented in books edited by Hopkins (1967) and Martin and Wright (1967). A collection of Quaternary studies presented at the 1978 AMQUA conference pertaining to the peopling of the New World was published by the Canadian Journal of Anthropology (Rutter and Schweger, eds. 1980).

3. LANDFORM ASSEMBLAGES

3.1 INTRODUCTION

Earth scientists rarely have the opportunity to deal with both the stratigraphic and morphological evidence for the events that they are investigating. Obviously, this luxury is only assumed to be available by those studying the most recent geomorphologic events as morphology is ephemeral. In fact, morphology is so ephemeral that many researchers (Richmond, 1965, Birkeland, *et al.*,1971; Miller, 1971, and others) have used the "freshness" or "youthfulness" of the landscape as a relative-age dating technique to discriminate between glacial events. Despite the cold, moist climate of the study area (see Section 1.2.6) the glacial landforms created by the last ice to have overriden the area are well preserved. They can be easily identified, even from aerial photographs, except in areas where post-Pleistocene erosion or deposition have altered them.

The regional nature of this investigation dictates that the major emphasis must be placed on the dominant, or principal, geomorphic agent rather than on the most recent identifiable event. A degree of generalization and subjectivity is unavoidable. An analogy may be drawn between the generalizations required in studying climatology from meterological data and the geomorphic generalizations of this study. The climatologist frequently makes generalizations which would not be acceptable at a higher resolution but nevertheless are accurate for the intended purpose. Lawson (1979) illustrated that only about 5 per cent of the sediments from the Matanuska Glacier were, in a strict sense, glacigenic in origin. If Lawson's (1979) criteria were used in a regional study such as this, it would be virtually impossible to create mapping units. The increased precision would, in fact, decrease one's understanding of the glacial history of the area. To illustrate this by returning to the analogy, it is well recognized that the instantaneous surface wind observations rarely reflect the pattern of the westerly circulation known to exist at these latitudes. It is the regional (westerly) pattern which is germane to any discussion concerning climates and climatic change.

In order to illustrate the implications of this philosophy for the present study, observe the following examples: A lateral moraine would be identified as such provided that the surface expression of the feature was clearly recognizable and the composition

of the deposit was dominantly of glacigenic origin. In every case the dominant mode of deposition will determine the map designation and there are no specific criteria which could be considered to be necessary and sufficient to include or preclude it from any landform unit. In most instances morphology provided the most significant clues for classifying the landforms and sedimentological evidence was usually employed to verify or augment the interpretation. The detailed stratigraphic evidence used by Lawson (1979) is useful to determine the specific depositional environment and, in some instances, these interpretations have been used in this study. The need for detailed stratigraphic studies is obvious and unending. It is hoped that this study will provide an incentive for additional studies to augment the information presented here.

The major premise here is that if there were repeated glaciations of the study area, then evidence for these events should be observable unless all manifestations of earlier events were destroyed by the most recent event. The excellent preservation of glacial landforms illustrates the suitability of this approach. In this section the landform units are described and the units are displayed spatially and topographically on the accompanying Landforms Map (see insert at back). The attempt to produce an explanatory model to encompass all of the features identified will be discussed in Chapter 6.

3.2 MAPPING UNITS

The landforms mapped in this study have been grouped into three quasi-chronologic categories; Pre-Pleistocene landforms, Pleistocene landforms and Holocene landforms. The Pleistocene and Holocene landforms are further subdivided according to the dominant geomorphic agent responsible for their formation.

3.3 PRE-PLEISTOCENE LANDFORMS

BEDROCK UPLANDS

All bedrock uplands have been classified as a single unit regardless of their lithologic composition. Bedrock uplands, as the name implies, occupy the regional topographic highs and are comprised of consolidated rock with associated veneers of talus and colluvium. The stratigraphic and lithologic units of these areas may be

determined from the Bedrock Geology Map (see Fig. 1.2).

3.4 PLEISTOCENE LANDFORMS

The Pleistocene forms are grouped according to their interpreted genesis. The landform units are again presented in a quasi-chronologic order. Landforms indicative of active ice are discussed first and features associated with ice wastage are discussed later. In order to simplify the Landforms map, and to expedite the interpretation of the related sedimentary environments, it was decided to choose the smallest practicable number of units. For example, thin mantles of lacustrine sediments or loess, that failed to produce significant differences in the landscape, were ignored for the purposes of mapping. Because of the large size of the study area the minimum area of mapped units was approximately two hectares, unless the landforms were considered to be particularly critical for the geomorphic interpretation. In such cases the sizes of the landforms, crevasse fillings for example, were exaggerated to allow their inclusion on the map.

3.4.1 GLACIAL LANDFORMS

MORAINE

The term *moraine* has been used very ambiguously throughout its history. Originally it was used to refer to ridges of debris at the margin of glaciers in the French Alps (Ritter, 1978). Some authors have used the term to indicate the depositional environment and the sedimentary characteristics of a glacial deposit, for example supraglacial morainic till (Boulton, 1980), but most authors have restricted the term to morphological features. In this study the term will denote depositional features whose form is indicative of ice-deposited sediments. A precise morphological definition is impossible because of the great variety of landform-producing environments beneath, and at the margins, of glaciers. Moreover, for the purpose of simplicity, moraines are considered to be deposited directly from glacier ice but obviously there are transitional forms where moraines grade into landforms reflecting a fluvial or lacustrine depositional environment. Till is therefore a common, but by no means the exclusive, sediment found

in moraines. The gradational forms such as kame moraines (Reimchen and Bayrock, 1977) and the ambiguous relationship between process and form, as in the transition between ice-moulded and ground moraine, renders a consistent and comprehensive classification system almost impossible to formulate.

The literature contains many attempts at classifying moraines. A classification based on the position of the landform relative to the glacier which formed it is part of a common vocabulary. Descriptors such as lateral, medial and end (or terminal) are commonly used in this system. Gravenor and Kupsch (1959) introduced a system which not only defined the moraines in respect to their relative position at the time of deposition, but also, using morphological evidence, they were able to group moraines into those that were deposited from ice which was active at, or immediately prior to the time of deposition, and those that were formed from relatively inactive ice. Prest (1969) expanded this system by sub-dividing the controlled deposition moraines into those that are oriented parallel to the direction of interpreted ice flow and those that are transverse to the flow direction. In addition, Prest's (1969) system recognizes subglacial moraines, ice-pressed moraines and ice-marginal moraines in each of the categories mentioned above. The advantage of Prest's (1969) system is that it provides genetic information pertinent to the glacier dynamics at the time of deposition which is germane to discussions of the geomorphic evolution of a region. The problem with this system is that if it is used at the mapping stage there is a considerable risk of generating circular arguments which, in turn, may lead to inappropriate conclusions. For example, on a vegetated surface it would be difficult to differentiate between Rogen moraines and washboard moraines which form subglacially, and push moraines which form at the ice front. From the perspective of a geomorphic interpretation, it would be advantageous to use a classification system such as Prest's (1969) as an intermediate step in the interpretation, and the field mapping should be conducted using a less-genetic nomenclature. It is for this reason that the landform classification system used for this study was adopted.

LATERAL MORAINES

Although they are poorly represented within the study area, some lateral moraine remnants were identified. An attempt was made to determine vestiges of ice-marginal positions using the criteria of off-set tributaries, notched spurs and breaks-of-slope as proposed by Chinn (1978). Notched spurs and breaks-of-slope proved useful for determining upper ice limits but no clear evidence of successive, ice marginal, still-stand positions was found. Along the sides of the main valley, where it transects the Foothills ridges, small remnants of lateral moraines were found. Remnants of lateral moraines were also found along the south side of the Brazeau River valley at the junction with the tributary Forestry Trunk Road valley. From the air a general break-of-slope may be observed along the valley between these points, indicating a possible ice-marginal position even though no moraines were found in the transitional area. The upper elevation of these features is about 1555 m at their western limit but this declines eastward to about 1463 m at the point where the moraines deflect into the tributary valley.

The morphology of these moraines is well preserved. Slope angles on the ice-proximal surfaces reach 34 degrees and on the distal side slope angles of about 10 degrees are common (see Fig. 3.1a). Postglacial stream cuts incised into the lateral moraines indicate that the associated till rests directly on local bedrock. The texture of the sediment is generally finer than that found in the pitted moraine units and most of the ground moraine units. No clasts with a-axes longer than 50 cm were observed. The clast shapes are mainly sub-angular to rounded and this indicates that they were probably fluvially modified at some time prior to their deposition by the glacier.

On the north side of the Brazeau River valley a smooth lateral moraine ridge extends downvalley from the southern end of the Cardinal Hills. The morphology is not well expressed. This may either be a result of post-depositional degradation or to limited initial moraine development. The second alternative is favored for two reasons. There are no extensive outwash trains or well-developed ice-marginal meltwater channels which would indicate a still-stand position for a significant period of time at this location and the valley broadens abruptly east of the Cardinal Hills and the ice would have expanded laterally beyond this point. The ice margin, therefore, would have fluctuated and distribute sediment over a broader area hence a decrease in the sediment supply to the glacier margin. The maximum proximal slope angle measured was 22 degrees while the distal

Fig. 3.1(a) Lateral Moraine, West Side of the Forestry Trunk Road Valley. Proximal Slope on Left.

Fig. 3.1(b) Lateral Moraine, North Side of Ruby Creek Valley. Proximal Slope on Left.

slopes of the ridge had angles of 11 degrees or less. A similar feature extends along the eastern side of the Brazeau River valley from the mountain front (Opabin Creek) to the Foothills ridges south of Thunder Lake (Coast Creek). All of the tributary valleys which enter the Brazeau River valley from the south appear to contain thick morainal deposits above an elevation of approximately 1620 m. Again it is suspected that the material which infilled these valleys, and now provides drainage divides in many of these structurally controlled valleys, was deposited from the margin of an active glacier located within the Brazeau River valley. From a genetic perspective, this would allow these deposits to be mapped as lateral moraine. However, from a geomorphic perspective it is important that the form of the deposits provide some of the evidence for the existence of an ice margin at this location before the term *lateral moraine* is applied to the landform. The deposits which infill these valleys fail to fulfil this criterion so the more conservative definition of hummocky moraine must be employed. If these features could be demonstrated to be lateral moraines, it would have important consequences for the alternative glacial model proposed in Chapter 6.

North of the Pembina River, along the east side of Lovett Ridge, a subdued lateral moraine extends around the ridge at an elevation of 1370 m. Slope angles are shallow, typically in the order of 9 - 12 degrees. The position and shape of this moraine suggests that deposition occurred from a lobe of ice which flowed out of the Foothills at this location but it is also possible that most of the material was supplied from the Athabasca Lobe. The hummocky and pitted moraines adjacent to, and topographically below, this unit indicate that during deglaciation in this location, the glacier degenerated from active to stagnant when the ice was still at least 60 m thick. This interpretation is based on the present topographic position of the upper surfaces of these features relative to the adjacent lowland areas where active ice forms and late-glacial/ post-glacial forms are common.

Lateral moraines and one end moraine were found along the eastern slopes of the Front Ranges (see Landforms map). The upper limit of these features follows a well defined break-of-slope located at an elevation of approximately 2200 m. The morphology of these features is relatively distinct but, because of their topographic position and relative absence of stabilizing forest cover, they are not as well preserved

as those along the eastern side of the Foothills. The ice-proximal slope angles are variable. Angles of 37 degrees have been measured but more commonly they are in the 13 - 20 degree range. The lithologies of the clasts in these moraines are all of local origin. Limestone and sandstone clasts dominate, with a few clasts of poorly lithified conglomerates. Clast shapes are sub-angular to angular with less than 10 per cent being rounded. The modal clast size was 15 cm but clasts with a-axes as large as 1.1 m were observed. Facetted surfaces were observed on limestone clasts but few striations, percussion marks or chatter marks were observed.

Lateral moraines are also associated with some of the small drainage basins that issue from the Front Ranges. The Neilson Creek and Ruby Creek valleys each contain well defined lateral moraines, in the form of nested ridge systems, which extend into the structural valley immediately east of the mountain front (see Fig. 3.1b). The pattern of the moraine ridges is very similar in both valleys. The lateral moraines are steep-sided, high ridges where they emerge from the mountain front and they converge relatively abruptly down the respective valleys. As the ridges converge, the slopes, height and size of the ridges tend to diminish. An extensive reconnaissance failed to identify associated end moraines, and the lateral moraines tend to grade into the ubiquitous till veneer which mantles this portion of the Foothills. It is worth noting that Reimchen and Bayrock (1977) note both lateral and terminal moraines within many of these valleys but these are not described nor are locations given. The disparity between this study and the Reimchen and Bayrock (1977) report must, therefore, remain unresolved.

Well preserved lateral moraines are found in the Cardinal River valley immediately east of the mountain front. The elevation of these moraines is about 1980 metres. The Cardinal Valley lateral moraines may be traced east as far as Red Cap Mountain, but beyond this point they grade into hummocky moraine and pitted moraine.

ICE-MOULDED MORAINE

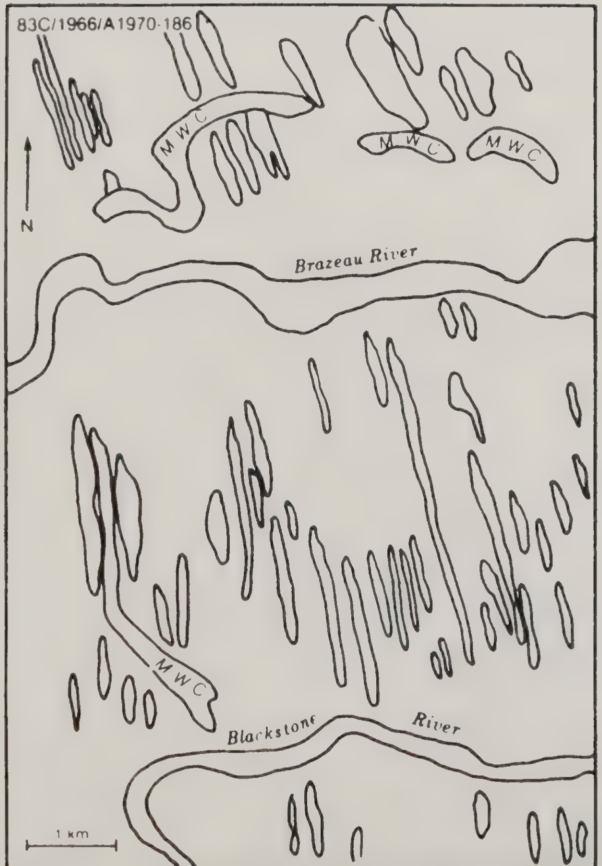
Both Boydell (1972) and Roed (1975) commented on the extensive swarms of ice-moulded landforms which are found within the major glacial valleys which issue from the Western Canadian Cordillera and occupy significant portions of the western plains immediately east of the Foothills. Gravenor (1953) provided similar information for the

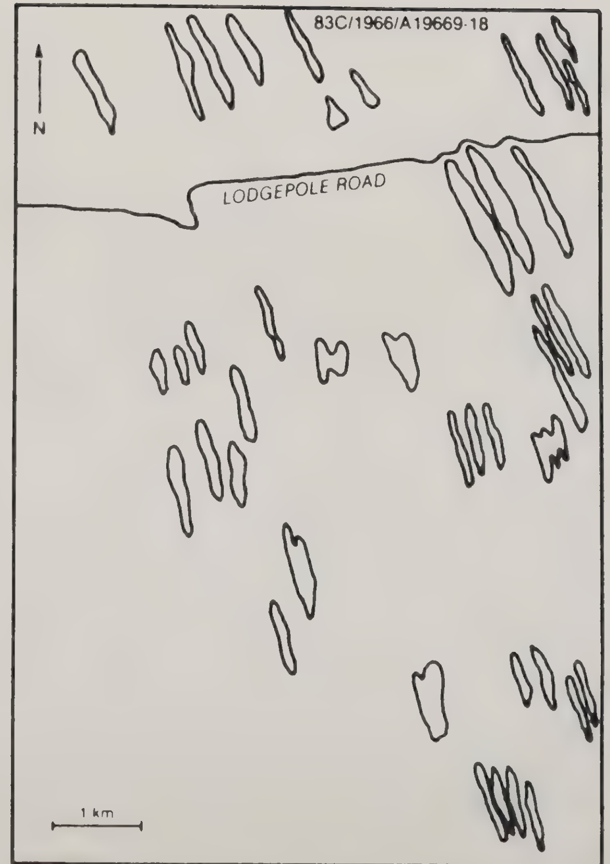
central and northern portions of Alberta. Despite the fact that no generally accepted explanation for the formation of ice-moulded terrain has been developed, they appear to be common features both in areas which were overridden by Laurentide ice and areas which were overridden by Cordilleran ice.

Menzies (1979) provides a comprehensive review of the literature pertaining to ice-moulded landforms (drumlins). It is clear that Hoppe's (1959) observations that ice-moulded landforms are oriented parallel to the direction of last ice movement remains generally accepted. Menzies (1979) illustrates that although considerable attention has been given to the geographic relationships between past ice margins and drumlin swarms, the evidence is still somewhat ambiguous. Similarly, no clear pattern of drumlin density, drumlin shape or drumlin size and/or complexity has thus far emerged. The strongest line of evidence they may provide is that they are useful ice-flow direction indicators.

More than 20 per cent of the study area is comprised of moraine that portrays distinct evidence of ice-moulding. Even a cursory look at aerial photographs reveals the extensive areas of drumlins and drumlinoid ridges which transect the study area east of the Foothills (see Figs. 3.2 and 3.3). Less conspicuous, but equally interesting, are the areas of transversely oriented ice-moulded features which are found within the Brazeau River valley near the confluence with the Southesk River (Fig. 3.4a) and east of the Pembina River, on the plains (Fig. 3.4b). Ice-moulded landforms probably provide the most unequivocal and easily observable evidence for the behavior of the ice flow across the study area.

To the east of the Foothills most of the study area, except for the uplands and postglacial valleys, is mantled with drumlinized moraine. General heights of drumlinoid ridges range between 5 and 20 m but relative relief of up to 45 m has been measured. The drumlin swarm can be traced continuously northwest into the Athabasca River valley which was evidently the source of the related ice. This is consistent with the map provided by Roed (1975). The ice-flow indicators on the Landforms Map and Fig. 3.3 indicate the orientation of the clearest and most representative examples of these drumlins. The general direction of flow is from the northwest (330 degrees True) but the bedrock uplands and preglacial valleys locally altered the flow direction. Along the


83C/1966/A1970-186
N
M W C
M W C
M W C
M W C
Brazeau River
Blackstone River
1 km


83C/1966/A19669-18
N
LODGEPOLE ROAD
1 km

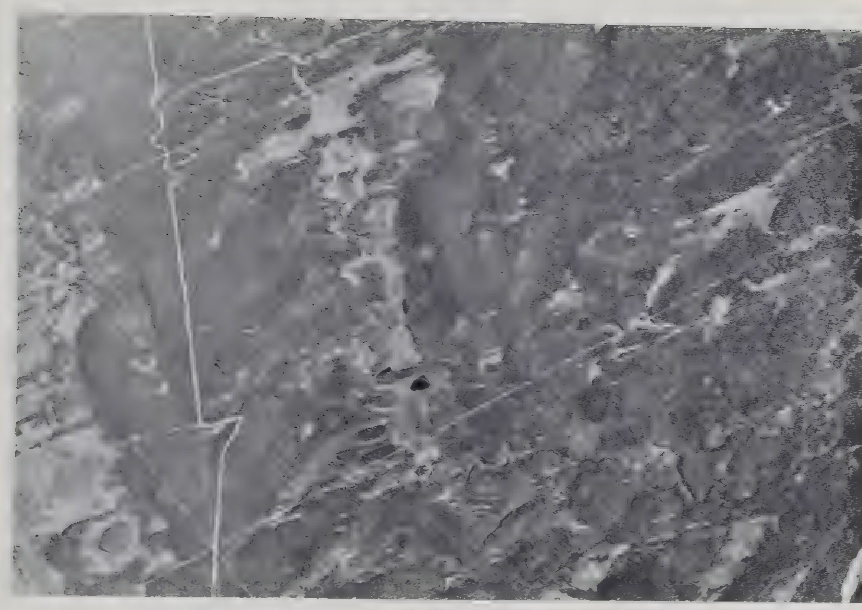


Fig. 3.2(a) Ice-Moulded Ridges on the Western Alberta Plains, east of the Foothills.

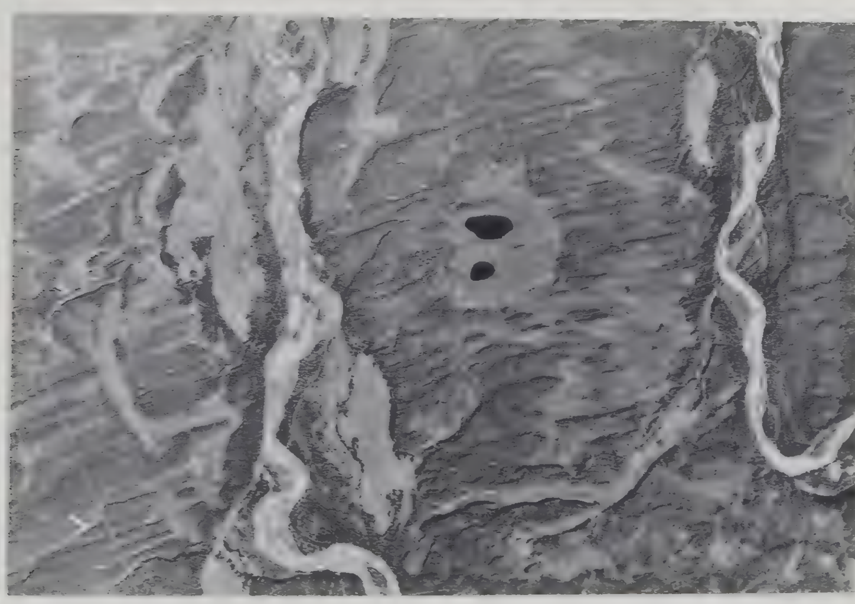


Fig. 3.2(b) Ice-Moulded Landforms on the Western Alberta Plains between the Brazeau and Blackstone rivers.

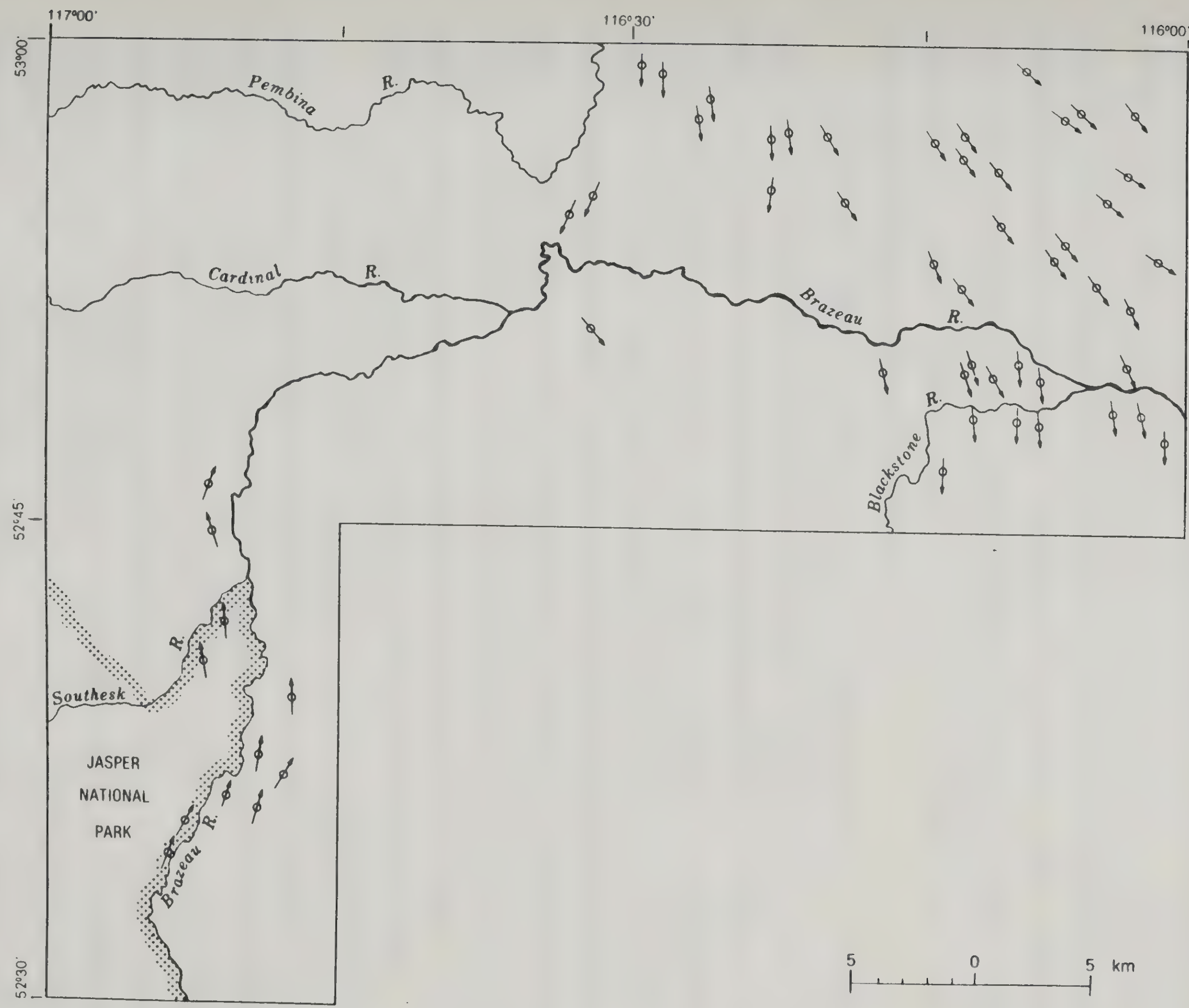


Fig. 3.3 Ice Flow Direction (inferred from selected drumlins)

eastern side of the map area the flow direction was from 300 degrees True whereas adjacent to the eastern margin of the Foothills a distributary lobe was deflected in toward the Foothills from 028 degrees True (see Fig. 3.3 and Landforms Map). The splayed pattern of the ice-moulded features implies that the Athabasca Lobe was expanding laterally as it flowed southeastward across the area. No ice-moulded whaleback (Flint, 1971) forms nor rock cores were observed within the study area. It is interesting to note that the stoss end of some of the drumlins sampled contained coarse sand with occasional rounded clasts up to 20 cm in diameter. Most drumlins are composed of a matrix-dominated mixture of large clasts in a sandy-silt matrix. No good exposures were found in these features therefore only the uppermost metre of sediment was observed in hand-excavated pits.

An additional point of importance is that the upland areas above an elevation of 1250 m have no drumlinized ridges across their upper surfaces. Instead, the ridges abut against the stoss side of the hills and re-form just below the break-of-slope on the lee side of the hills. The upper surfaces of the uplands are veneered with ground moraine, and the limited available borehole data suggest that approximately 6 m of overburden covers the bedrock. However, no ice-moulding of these sediments was observed. A paucity of unconsolidated material cannot therefore be cited as an explanation for the lack of ice-moulded features in these areas. Perhaps there is a critical ice thickness required to initiate the glacial processes that result in drumlins and flutings or there is some critical basal condition such as water (Shaw, 1983) which was not available.

Within the valleys of the Foothills there are two suites of ice-moulded landforms, both of which occupy the bottoms of tributary valleys, well above the floor of the main Brazeau River valley. One suite lies below the lateral moraine which wraps around into the Forestry Trunk Road valley. The floor of the tributary valley is at an elevation of approximately 1400 m, about 90 m higher than the adjacent floor of the main valley. The ice surface gradient indicated by this suite was southeastward out of the Brazeau Valley and into the Forestry Trunk Road valley. The deflection of ice into this valley implies that some obstruction must have been present east of the Foothills, as local topography would otherwise dictate a flow toward the northeast. A meltwater channel associated with this suite of landforms (discussed in Section 3.4.2) indicates that meltwater flowed

across the foothills ridges from west to east at an elevation of approximately 1380 m shortly after the lateral moraine was formed.

Further west a small swarm of drumlinized moraine forms indicate that ice flowed out of the main valley into the Crooked Creek valley. Again, it appears that these landforms are immediately below a lateral moraine (?) which clearly shows a deflection of ice from the Brazeau valley into the Crooked Creek valley. Unfortunately, the lateral moraine designation for this landform has not been demonstrated conclusively and it has been mapped as a hummocky moraine. The elevation of the tributary valley threshold is at approximately 1400 m and deflection into this valley would occur only if an obstruction prevented a downstream flow along the main valley. In order for a distributary lobe of ice to flow down this valley the ice surface gradient must have been toward the north (i.e toward the Pembina River valley). The entire Pembina River headwaters lie within the Foothills, therefore any ice accumulation in this area would have been of local origin.

A large area of drumlinized moraine occupies the floor of the main valley where the Brazeau River exits from the mountain front. The orientations of the ridges are sub-parallel to the axis of the valley. The upper surface of the drumlinized moraine is dissected by meltwater channels, particularly between the Southesk and Brazeau Rivers. Postglacial streams have incised the glacial deposits producing the best stratigraphic exposures within the study area. Forty metres of till occur along the Brazeau River and up to 28 metres along the Southesk River (see Chapter 4).

The morphology of the drumlinized moraine in this area is significantly different from the drumlins on the Western Alberta Plains. On the plains the ridges tend to be attenuated and relatively long whereas the drumlins at the mountain front have a smaller length-width (L/W) ratio (see Table 3.1). They are best expressed just downvalley from the mountain front and their form becomes less distinct with distance downvalley. In fact, in that direction, they begin to resemble the crested and incomplete drumlins described by Lundqvist (1970). Near the junction of the Southesk and Brazeau Rivers an area of wavy-transverse ridges exists with the curves or horns of these features deflecting downvalley in a manner reminiscent of Rogen moraines described by Lundqvist (1969). Downvalley of the Southesk River, the drumlins are again poorly formed, with many complex features which appear to comprise two or more drumlins in mutual contact.

These complex drumlins were not superimposed forms but rather conjugate forms similar to those described in Eastern Cumbria by Rose and Letzer (1977). Near the mouth of Thistle Creek transverse ice-moulded features again become more prevalent than longitudinal ones.

Within the mountains to the west of the study area a swarm of ice-moulded landforms exists downvalley of Brazeau Lake. Morphologically, these appear to be classic drumlins but recent stream dissection has revealed that they are composed entirely of quartzitic bedrock (Gog Group according to the Geologic Map of Alberta, Alberta Research Council, 1972) with a thin veneer of colluvium. This bedrock appears to be *in situ* and undistorted by glacial action. Unlike the other ice-moulded features discussed in this study these are thought to have been formed entirely by glacial erosion even though no striations or clear evidence of glacially polished surfaces were observed.

INTERPRETATION AND DISCUSSION OF ICE-MOULDED MORAINE

Figure 3.3 illustrates the general ice flow pattern of Athabasca Lobe ice across the study area. The splayed pattern of the central portion gives way, south of the Blackstone River, to flow indicators trending north-south. These show that ice moved obliquely upslope toward the Foothills in this zone. Such a flow may be explained by the following model: Laurentide ice existed to the east of the study area and provided a barrier to Cordilleran ice flowing eastward. Roed (1968,1975), Boydell (1972,1978), Stalker (1977) and Peters (in press) provide evidence to support this hypothesis. If the Athabasca ice was simply flowing along a corridor between the Foothills and Laurentide ice, it is difficult to explain why the local ice flow indicators show lateral expansion north of the Brazeau River valley but south of the Blackstone River the flow was oriented to the southwest, towards the Foothills. It is possible that a lobe of Laurentide ice locally bulged toward the Foothills causing greater compressive force in this area. Alternatively, a lobe of ice extending from the Brazeau River valley to at least the eastern margin of the Foothills may have deflected the Athabasca Lobe in the immediate vicinity of the Brazeau Valley. Down-flow of the confluence of the two ice lobes, the Athabasca glacier would have been freed of the lateral constraint and would have therefore expanded back toward the Foothills. It is worth noting that if this explanation is correct, the Brazeau lobe

83C/1966/A19668-87
N
Southesk River
Brazeau River
MELTWATER CHANNEL
1 km

Fig. 3.4(a) Transversely-oriented ice-moulded landforms above the confluence of the Southesk and Brazeau rivers.

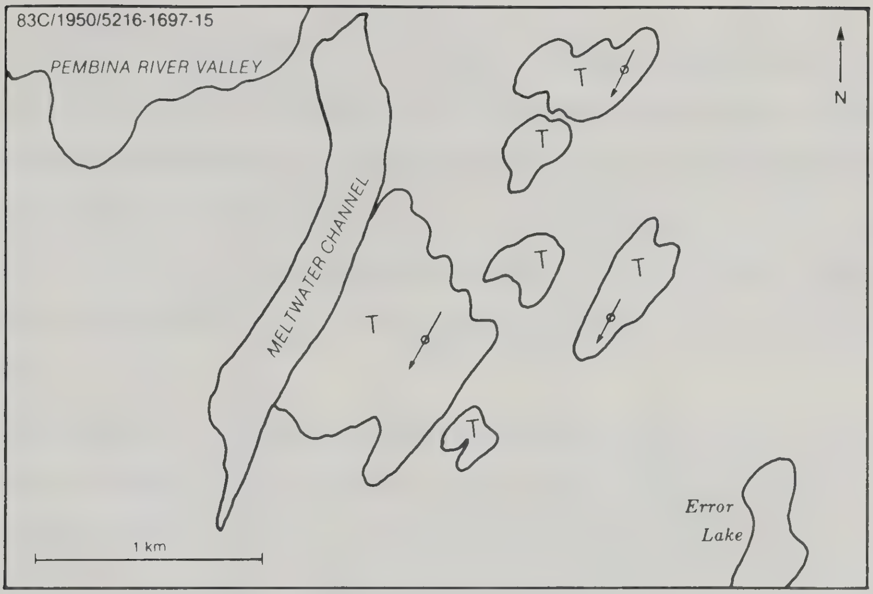


Fig. 3.4(b) Transversely-oriented ice-moulded landforms along the Pembina River, immediately east of the Foothills.

did not issue much beyond the eastern edge of the Foothills at the time of formation of the ice-moulded moraine. The observed orientations of two small ice-moulded moraine features along the north side of the Brazeau River valley within the Foothills indicates that Brazeau ice probably flowed north, into the Crooked Creek valley. This suggests that at the time of formation of the drumlinized moraine, and the hummocky (lateral ?) moraine, the ice level in the Brazeau Valley was significantly higher than in the Pembina Valley to the north.

The paucity of evidence of glacial erosion (cirques, troughs, aretes, etc.) within the Foothills, except for the major river valleys which transect the Foothills, combined with the evidence provided by the ice-moulded moraine, indicates that the flow of Athabasca ice in this area was restricted to the area east of the Foothills. The explanation invoking a Brazeau Lobe of ice is supported by the lateral moraine on the southeast flank of the Cardinal Hills and the extensive esker complex that can be traced discontinuously to the eastern edge of the Foothills along the Brazeau River valley (see Landforms Map). In addition, the lack of a continuous suite of landforms from the Athabasca Lobe into the Brazeau Valley from the east indicates that some obstruction must have prevented such a flow, as regional topography would not prevent it.

If the second explanation, which is consistent with the geomorphic evidence, is valid it suggests two things:

(1) ice from the Brazeau Valley abutted against, but only locally deflected, the flow of Athabasca ice along the eastern margin of the Foothills, and,

(2) the Foothills to the south of the Brazeau River valley were not contributing significant amounts of ice to the plains area further east.

The interpretation of glacial processes responsible for the formation of ice-moulded landforms is difficult because of their variability in composition and morphology. Menzies (1979) illustrates the fact that features displaying the drumlinoid shape have been found to contain every conceivable combination of sediments from laminated silt and clay, through stratified sand and gravel to compact, clay-rich till and bedrock. Gravenor and Meneley (1958) investigated the geographic relationships between drumlins and other glacial landforms. They found that drumlins tend to occur in low-lying areas whereas ice-moulded features are characteristic of steeper slopes and upland

areas. Gambier (1984) illustrated some of the complex sedimentology associated with drumlins in the foothills of the Athabasca Valley. To date no single explanation of drumlin formation appears to adequately explain their distribution. In fact, recent studies (Shaw, 1983; Shaw and Kvill, 1984) question even such accepted facts as the need to postulate active ice at the time of their formation. The evidence from this study does, however, allow the dismissal of some of the proposed models of drumlin formation for the swarms identified in this study.

The Gravenor (1953) model of differential deposition of till near a terminal moraine fails to account for the rock drumlins found upstream (within the First Ranges) of the drumlins comprised of glacial sediments. Perhaps this is not important because the rock drumlins could well be similar in morphology alone and may have been created by an entirely different process. Nevertheless, this model predicts a narrow band of drumlins near the terminal position of the lobe. The drumlin swarm from the Athabasca Lobe may be traced from the Athabasca Valley to the Rocky Mountain House area; a distance of more than 180 kilometres. This can hardly be construed as a 'narrow band' developed near the former terminus of the Athabasca Lobe.

The notion that bedrock nuclei are a necessary pre-condition for drumlin formation (Hoppe and Schytt, 1953; Gillberg, 1976) seems unlikely as the friable sandstone and fissile shale underlying the drumlinized moraine would not likely withstand the shear stress and corrasion at the sole of a warm-based glacier.

Alden (1918) noted that the drumlin swarms in southwest Wisconsin form a fan-shaped pattern which suggests radial extension of the glacier. It is interesting to note that Figure 3.3 in this study and Figure 9 in Roed (1975) reflect a similar pattern. Shaw (1980a) used these drumlins which emerge from the Athabasca Valley to illustrate a model of drumlin and fluting formation based on secondary flow within the ice. Shaw (1980a) explained the secondary flow as a consequence of transverse folding of a glacier produced by lateral compression and longitudinal extension. The evidence presented in Figure 3.3 is consistent with Shaw's (1980a) hypothesis. The ice-flow pattern illustrated by the mapped forms appears to reflect a lateral constraint to the east of the study area, and the drumlin morphologies are best expressed in areas immediately adjacent to major valleys which transect the Foothills (e.g. the Brazeau River valley). A

significant difference between Alden's (1918) model and that proposed by Shaw (1980a) is that Alden requires radial expansion of the glacier in order to produce splay crevasses at the base of the ice whereas Shaw's (1980) model requires transverse compression to produce folds in the ice. The fan-shaped pattern supports Alden's (1918) model but the evidence of ice-moulded moraine ridges having been deflected back toward the Foothills instead of radiating out onto the plains argues well for Shaw's (1980) laterally-confined flow model. It is likely that this difference will remain enigmatic at least until a more detailed study of the orientation and spatial distribution of the entire drumlin swarm is undertaken. In addition, the fact that many drumlinoid features are similarly oriented on both the proximal and distal sides of the upland areas, with limited deflection, suggests that the ice must have been sufficiently thick to mask the effects of these positive relief features during the phase(s) of drumlin formation

Some researchers (Reed *et al.*, 1962; Muller, 1963, 1974; Vernon, 1966) postulate that former ice dynamics can be inferred from drumlin swarms through analysis of the spacing between drumlins and of their length-width ratios. Lundqvist (1969) recognized a transition of forms of ice-moulded moraine from Rogen moraine, to separate crecentic ridges, to complete drumlinoid forms.

Embleton and King (1975, p. 407 and 420-421) used evidence, presented by Reed *et al.*, (1962) from central New York State and Hollingsworth (1931) from the Lake District of England to illustrate that drumlins may become more elongated and streamlined where the ice is thicker and moving faster. If this is so, the drumlins in the northeast section of this study area were deposited under the fastest flowing and thickest ice and those at the mountain front, within the Brazeau River valley, were formed under the thinnest and least active ice. Measurements of the length-width ratios of drumlins in the study area revealed the information depicted in Table 3.1. This hypothesis would be consistent with the interpretations discussed in Chapter 6. It implies that the major ice lobe within the study area was the Athabasca Lobe and the thickest and most active ice was well to the east of the Foothills. In contrast, the Brazeau Lobe was probably thinner and less active.

Recently, Moran *et al.*, (1982) interpreted the ice-moulded landforms of the prairie region of North America as being the product of ice-thrust erosion at the base of

LOCATION	L/W	RANGE
Rt. side of Brazeau River, east of mtn. front.	2.7:1	4.3:1 - 2.0:1
Left side of Brazeau, upstream of Thistle Creek.	3.7:1	6.3:1 - 2.5:1
North of Brazeau Valley, distal side of uplands.	7.1:1	11.5:1 - 5.3:1
Proximal (N) side of uplands.	10.7:1	19.3:1 - 4.5:1
Between Blackstone and Brazeau Rivers.	12.9.1	23.0:1 - 9.0:1
Along Elk River, N.E. corner of study area.	19.8:1	23.1:1 - 14.0:1

Table 3.1 Length/width ratios of selected drumlins and flutings.
(A minimum of 10 features were measured at each location).

the glacier followed (in the lee of the thrust masses) by deposition of the quarried material and subsequent sub-glacial moulding of the thrusted blocks. In this model the glacier is cold-based at the time of ice thrusting and subsequent streamlining, although warm-based conditions must exist a short distance up-flow in order to generate the porewater pressures required for the quarrying of the thrust blocks. This model, although useful to explain the distribution and internal composition of drumlins on the prairies, is not easily adapted to explain the features mapped within this study area for the following reasons:

(1) There are no obvious depressions, which Moran *et al.*, (1982) claim are normally found associated with these features, within the mountain valleys upstream of the drumlin swarm found immediately east of the mountain front. The valley shape is very smooth and appears to have been formed from corrasion more than quarrying. Moreover, the valley bottom contains only a thin layer of late-glacial and post-glacial deposits over bedrock which dips steeply toward the southwest. This would provide a poor environment for the ice-thrust erosion necessary for drumlin

formation according to this model.

(2) The drumlins produced by the Athabasca Lobe can be traced from the Athabasca Valley across the eastern portion of study area, and southeast nearly as far as Rocky Mountain House. If these drumlins developed according to the conditions described by Moran *et al.,* (1982), they must have been formed time-transgressively insofar as that study suggests the necessary basal ice conditions would occur in a geographically restricted zone of perhaps 2 km in width.

(3) Because the necessary basal conditions postulated by Moran *et al.,*(1982) for thrusting are thought to be restricted to a specific portion of the glacier, the observed suite of ice-moulded landforms which extend from the Athabasca Valley to Rocky Mountain House could have only formed time-transgressively. This would require a retreating ice front with active ice up-glacier of the terminal margin. This is incompatible with the landforms, such as eskers, transverse moraines and crevasse fillings which have been mapped throughout the area.

(4) Ice-thrust erosion may have acted in some areas, as the stoss, or northern, slopes of the Tertiary Uplands appear to have experienced a form of erosion which removed large rectangular blocks. The small drumlins which are found inside the basins of the quarried areas would be difficult to explain by the Moran *et al.,* (1982) model unless two or more episodes of ice-thrust erosion were postulated.

Support for the model of Moran *et al.,* (1982) lies in the observation that the drumlinized terrain invariably overlies poorly consolidated permeable bedrock of the Alberta Group, the Brazeau Formation or the Paskapoo Formation.

Regardless of which of the many models of drumlin formation one accepts, it is probably safe to assume that these features provide a good indication of the direction of the last ice movement and they were formed subglacially beneath a non-stagnant ice body. The degree of activity in the glacier remains controversial although the kettle in the center of Fig. 3.2b with ice-moulded features grading into and out of it does suggest that the ice must have been relatively inactive at the time of drumlin formation.

The distribution of ice-moulded moraine and the orientation of these features may provide clues as to the relationship that existed between the Laurentide and

Cordilleran glaciers during the time when they coalesced. Unfortunately the antecedent conditions for the formation of ice-moulded moraine are not known, but if lateral compression and/or longitudinal extension of a glacier correlate with the formation of drumlins and flutings, as Shaw (1980a) suggests, then during the time when the ice-moulded moraine was forming along the eastern slopes of the Foothills, the Cordilleran ice must have been confined between the topographic barrier provided by the Foothills and the Laurentide glacier. The absence of similar landforms to the east of the contact zone suggests that a similar subglacial environment was not found there. The orientation of the ice-moulded moraine provides additional grounds for speculating that perhaps the Cordilleran ice retreated faster or earlier than did the Laurentide glacier. Field observations described in the previous section, the evidence provided by Figure 3.3 and the Landforms Map document the fact that the ice-moulded moraine ridges trend inward toward the Foothills south of the Brazeau River valley rather than away from the Foothills as the regional topography would dictate. This would occur if the Laurentide glacier was to dominate the Cordilleran glacier. A Laurentide re-advance could produce such a flow direction, or, if the Cordilleran glacier declined before the Laurentide glacier, the same flow pattern would be produced. To date there have been no studies from southwestern Alberta which indicate that the Laurentide glacier fluctuated while the Cordilleran glacier remained relatively static; in fact, the reverse may be true (Boydell, 1978). Therefore, there is some merit in postulating that the orientation of the ice-moulded moraine indicates that the Cordilleran glacier receded earlier, or faster than the contiguous portion of the Laurentide glacier. Further investigation of this hypothesis must necessarily await a clearer understanding of the processes responsible for the formation of ice-moulded moraine.

TRANSVERSE ICE-MOULDED LANDFORMS

Several examples of streamlined, transversely-oriented features which appear to be genetically related to drumlins were noted. These warrant attention as they have never been documented or described before in the Foothills of Alberta. They appear to be comprised of till which is similar in texture to that found in adjacent drumlin forms and no boulders were observed to protrude above the surface of these features.

Hand-textured samples from pits approximately 1 metre deep suggest that the percentage of silt and clay may be slightly greater in the transverse features but larger exposures and more detailed sampling would be required in order to verify this tentative observation. The streamlined cross-sectional shape of these features, together with the transition from transverse forms to classical drumlin shapes, provides the basis for the term transversely-oriented ice-moulded moraine. They appear to be very similar to those described by Lundqvist (1969) and Aario (1977) in Scandinavia. Lundqvist (1969) suggested that Rogen moraines are genetically similar to other forms of ice-moulded moraine but they seem to occur in areas where the landscape forms a broad depression. Minell (1977) described sedimentary evidence to support an interpretation that transverse moraines form in areas of compressive flow, possibly due to thrusting caused by frost adhesion between the subglacial sediment and the sole of the glacier. Shaw (1979) provided additional support for this model by describing the stratigraphy in greater detail, particularly in terms of till fabrics and fold structures within the sediments. A morphological observation noted by Shaw (1979) was a series of escarpments on the distal side of these features. Perhaps these small-scale structures have not been preserved in the Alberta Foothills, but none were observed in the features mapped in this study. Shaw (1979) generally agrees with Lundqvist (1969) and Minell (1977) that Rogen moraines are produced by compressive flow at the base of the glacier but he is more specific and, at least for the examples he studied in Sweden, he feels that the deformation may have been englacial and deposition may have been due to subsequent meltout. For the transverse features within this study, it is safe to assume that they formed in areas which were subjected to longitudinal compression. The compression may have been caused by the ice traversing an over-deepened portion of the valley or by a lateral constriction in the valley. This interpretation is supported by the associated landform evidence associated with the transverse ice-moulded features found to the east of the Pembina River. Classic Rogen moraines are also found immediately south of the Southesk River, approximately 3 km southwest of its confluence with the Brazeau River. Here Lundqvist's (1969) model is less appropriate as the valley floor is not longitudinally concave. Rather, the width of the valley decreases abruptly immediately downvalley from this location. Similarly, further up the Brazeau River valley, crested

drumlins, incomplete drumlins and transversely-linked drumlins are found along the right-hand side of the valley near the mouth of Opabin Creek. The floor of the main valley is not longitudinally concave but the ice-moulded landforms indicate that the glacier expanded laterally into a piedmont lobe in this area. This is consistent with the refined compressive-flow models offered by Minell (1977) and Shaw (1979). Such factors as a concave longitudinal profile, valley constriction (Haldorsen and Shaw, 1982), or lateral expansion of the ice into a piedmont lobe, would all tend to produce longitudinally compressive flow. These factors all correlate with transverse ice-moulded landforms of the study area.

GROUND MORAINE

Areas mapped as ground moraine are mantled with glacigenic sediments of varying thickness which largely mask the pre-existent topography. The ground moraine surface is usually undulating but evidence of structures longitudinal or transverse to the interpreted flow direction, ridges, kettles or steep-sided hummocks, are absent. Typically, the deposits which form ground moraine are thinner than those which comprise hummocky or pitted moraine and they are usually located in topographically low areas. Measurements taken from Grave Flats and similar, but unnamed, valleys north of the Brazeau River indicate a modal sediment thickness in the order of 2-5 m but the lack of exposures in other areas precludes the presentation of a general range of the thickness of ground moraine deposits within the Foothills. Drill log records from the portion of the study area to the east of the Foothills indicate that 3 m is perhaps the modal thickness of till and no ground moraine thickness in excess of 10 m was observed. Nobles and Weertman (1971) suggest that the thickest till deposits should be located in the topographic depressions but within this study area these areas usually contain surface water and/or thick mantles of organic terrain, hence it is possible that the maximum till thickness could exceed the value presented above.

Till is, of course, the most common glacigenic sediment found in ground moraine. The sedimentary characteristics of the till comprising the ground moraines are highly variable. In the lower elevations of the tributary valleys within the Foothills the till is typically of local origin and dark brown or dark grey in color. This coloration is almost

certainly due to the dark shale and coal which have been incorporated into it. Clasts are dominantly of calcareous or siliceous sedimentary rocks and angular to sub-angular in shape, although well-rounded clasts are present at most sites. A brief description of the characteristics of these deposits is presented in Chapter 4. In general, they are not a single massive unit but rather they tend to be loose and rubbly except where indurated by carbonate precipitates. Striations were observed on some of the carbonate clasts but neither striations nor facetted clasts are common. Although most clasts display some evidence of having been fluvially rounded, fractured surfaces and broken clasts were evident at most exposures. Inclusions of poor-to-well sorted sediment were observed in most of the drilling logs and exposures. Usually these inclusions of sorted material displayed secondary stuctures such as folding and faulting. Usually there was little evidence of bedding within the sorted inclusions and the sorted facies commonly graded into heterogeneous deposits.

Ground moraine is polygenetic and therefore not particularly revealing of the geomorphic history of the area. It may consist of a layer of lodgement till, with only an irregular mantle of ablation till, or it may consist of multiple till units, including basally deposited material, englacially transported material and ablation deposits, as illustrated by Sugden and John (1976, Fig. 11.1, p.214). Goldthwait (1971) argued from empirical evidence that rock debris is usually concentrated in the bottom few metres of a glacier (20 to 100 feet; Goldthwait, 1971 p. 9). Subsequent deposition may be by lodgement (Chamberlain, 1894), meltout (Goodchild, 1875) or as supraglacial debris which slides, falls or flows off the ice surface (Boulton, 1971). Boulton (1976), Dreimanis (1976), Shaw (1977) and Lawson (1979) have attempted to formulate genetic classification systems for tills. Each of these systems requires that the very detailed sedimentary characteristics of the deposits be logged in order to classify the till unit. As desirable as this evidence may be to determine the genesis of the material, it is beyond the scope of a regional study. Few good exposures of ground moraines are available in the study area, and because the associated landforms usually occupy areas of relatively subdued relief, they are frequently masked by organic deposits, loess or a heavy forest cover. A program of drilling or excavation would be required to reveal sub-surface stratigraphic characteristics which might furnish valuable evidence such as thrust or sheared

structures, preferred clast orientations, or compaction from which a genetic interpretation of the tills could be determined.

There are, however, useful geomorphic indicators of the genesis of some of the ground moraines. A mantle of till was observed draped over the sediments of the esker complexes. If the glaciofluvial deposits were developed englacially or subglacially then, by the law of superposition, the till must have been englacial or supraglacial in origin. Similarly, near the eastern edge of the study area, several large erratics were observed. These features project more than a metre above the present ground moraine surface and an unknown distance beneath the surface. It would seem most probable that these large boulders were released onto the surface from a wasting glacier. Whether the associated ground moraine was deposited simultaneously or prior to this event is not known.

From these few observations, it appears that at least some of the ground moraines were deposited from englacially or supraglacially transported debris. Boulton and Eyles (1978) describe an environment where coarse-textured till mantles the underlying landforms, without providing any pronounced relief features, as being one in which rapid frontal retreat of the glacier occurred. Meltwater channel sediments, crevasse fillings and collections of surface boulders are predicted by their model, all of which are evident in the area east of the Foothills.

Two sub-categories of ground moraines were mapped; moraine-mantled bedrock and lacustrine-mantled ground moraines. Moraine-mantled bedrock is a unit gradational between ground moraine and exposed bedrock and/or colluvium. Chinn (1978) noted similar moraine forms in the Southern Alps of New Zealand. In most locations a large amount of local colluvium is incorporated into this unit, particularly on steeper slopes. Because much of the till is of local origin it is difficult, without detailed sampling, to subdivide these into separate units. Undoubtedly these deposits were reworked contemporaneously, or subsequent to deposition, and contemporary geomorphic processes are continuing to modify and eradicate these landforms. The till found within the moraine-mantled bedrock unit is usually indistinguishable from that of the adjacent ground moraine.

Lacustrine-mantled ground moraine is common within some of the structural valleys in the Foothills and along the eastern slopes of the Foothills. Typically these are

areas of subdued relief and, in depressional sites, are mantled by organic deposits. As discussed in section 4.2.2 (B9/R) some exposures display a transition from till to lacustrine sediments. This, together with the topographic evidence, indicates that these landforms were created by deposition into ice-marginal ponds. The sandy-silty texture of the lacustrine sediments suggests that there must have been some flow within the ponds to remove much of the clay fraction which one would expect to be the product of glacigenic sediments formed from a shale-rich parent material. Fine sand and coarse silt were, by far, the most common textures observed in these deposits. If it were not for the occasional larger clast (dropstone ?) one might confuse these with aeolian sediments.

Although Reimchen and Bayrock (1977) mapped most of the Foothills, except in close proximity to the major river valleys, as having a mantle of colluvium (because of the predominance of local lithologies within the deposits and the rounded but unabraded appearance of the Foothills ridges) erratics can be found throughout the area. Most commonly these erratics are limestone or sandstone but quartzites were also found by this author. Invariably the erratics comprise a very small fraction of the clasts present. Local streams and gullies proved to be the most likely spots to find erratics. Although not ubiquitous, erratics may be found to elevations of at least 2075 metres. In fact, except for the ridges of exposed bedrock and areas that have obviously been subjected to mass-movement processes, none of the Foothills ridges was found to be free of erratics. This indicates that, despite the unglaciated appearance of the Foothills, ice covered these areas during at least one glaciation.

HUMMOCKY MORAINE

The term hummocky moraine is purely a descriptive term which has been used very ambiguously in the literature. It pertains to a variety of landforms which are composed of glacigenic sediments and display an undulating, chaotic topography. The term has been used to describe knob-and-kettle topography with a range of forms which includes sharply-pointed ridges and hillocks and gently rounded mounds (Lundqvist, 1943 and cited in Hoppe, 1952), moraine plateaux, closed linear ridges and

ice-walled channels (Gravenor and Kupsch, 1959), and small, circular or oval ridges (Hazelton, 1968). Sediments described by other authors working in hummocky moraine areas include compact till (Hoppe, 1952), stratified silts and clays (Christiansen, 1956 and Mathews, 1963), and a complex assemblage of dominantly clay-rich till with subordinate inclusions of sand and gravel and a mantle of lacustrine silts and clays or "washed" till (Gravenor and Kupsch, 1959). Jennings (1984) provides a summary of the theories of formation of hummocky moraine. He has identified five types of theories. Three of the five mechanisms he outlines are:

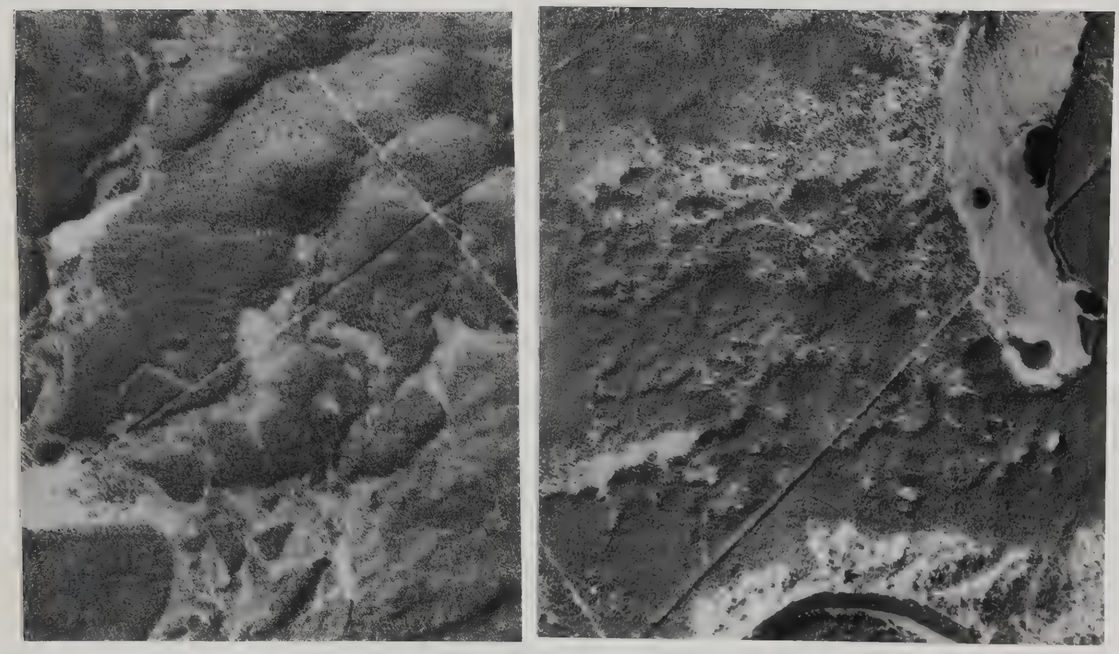
(1) The let-down theory (Gravenor and Bayrock, 1955) wherein a mantle of supraglacial sediment is lowered differentially onto the surface during the ablation of a stagnant glacier.

(2) The ice-press theory (Hoppe, 1952 and Stalker, 1960b) which explains hummocky moraines as the product of subglacial extrusion of saturated, unfrozen basal till into crevasses and cavities beneath the ice.

(3) The active ice theory which encompasses a large and diverse group of studies that attempt to provide explanations for Rogen moraines, Veikki moraines and ribbed moraines. These forms display sedimentary characteristics and morphologies indicative of a genesis beneath an active ice body.

The periglacial and ice-diapirism theories presented by Jennings (1983) pertain to only a subset of hummocky moraine which includes circular, semi-circular or elliptical ridges. These do not appear to be useful for landforms found in this study.

In this study hummocky moraines were distinguished on the basis of morphology. Figure 3.5 illustrates the morphological differences between pitted moraines and hummocky moraines. Hummocky moraines consist of rounded, randomly distributed mounds and depressions. The sediments comprising these features are typically poorly consolidated, unstratified deposits of cobbles and boulders in a matrix of sand and silt. Inclusions of sand, gravel and occasionally silt and clay were observed. Only in a few sites where silts and clays were at the surface in topographic depressions were the finer materials stratified and sub-horizontally bedded. Typically the sorted inclusions of sub-gravel fractions displayed evidence of considerable deformation. It is assumed that these disturbances occurred during the final deposition of the material from the glacier.



Hummocky Moraine

Pitted Moraine

Fig. 3.5 Morphological comparison between hummocky and pitted moraine

Occasionally, for example in the tract of hummocky moraine immediately upvalley of the confluence of the Cardinal and Brazeau rivers, large limestone erratics are embedded in these deposits but extend one or two metres above the adjacent moraine surface. The subsurface extent of these boulders was not determined but it is fairly clear that they are contained within, and not resting on top, of the surface of the moraine. If this observation is correct, the origin of at least part of this landscape was ice-marginal rather than subglacial. This would be consistent with the spatial distribution of these forms within the study area.

Local relief may be in the order of 10 m but in no instance does it reach the 50 – 70 m range of the hummocky moraines in central Alberta (Gravenor and Kupsch, 1959 and Stalker, 1960). Slopes are more subdued than in the kettles of the pitted moraine, with slope angles typically less than 25 degrees. Organic deposits frequently occur in the

depressions and both occupied and abandoned stream channels dissect the moraine surface. The relatively fine texture of the tills, particularly in the foothills where large amounts of local shale were incorporated, inhibits subsurface drainage. This has stimulated the development of muskegs and fens in depressional sites. Poorly developed proglacial drainage systems and ice-marginal pond deposits are frequently found in association with hummocky moraines.

Although there are few good exposures from which to accurately determine the depths of deposits associated with these landforms, they appear to be generally thicker than 5 m and frequently thicker than 10 metres. Therefore, hummocky moraines may be tentatively differentiated from ground moraines and moraine-mantled bedrock on the basis of the thickness of the deposit as well as on morphologic and sedimentary criteria.

Hummocky moraines flank the Brazeau River and Cardinal River valleys. In addition, they occupy a part of the interfluve between the Cardinal and Brazeau Rivers at an elevation of approximately 1400 meters. The relative positions of these landforms are comparable to those of lateral moraines. They differ from lateral moraines morphologically in that there are no significant ridges sub-parallel to the direction of ice flow, or clearly observable boundaries separating these from adjacent deposits. Commonly hummocky moraines grade into pitted moraines or moraine-mantled bedrock (see Landforms Map). The main geomorphic evidence to support the view that they are ice-marginal features is that they create unusual divides in tributary valleys and differentially infill structural valleys which lie transverse to the interpreted direction of ice flow down the major valleys. This evidence is particularly will expressed in the minor structural valleys along the southern rim of the Brazeau River valley.

The interpretive import of the hummocky moraines mapped in this study is diminished by the fact that the form and sedimentary characteristics of these are sufficiently ambiguous to preclude a specific genetic interpretation. The nomenclature adopted herein is deliberately vague to allow the classification scheme to reflect the relative confidence that can be expressed concerning the genesis of these landform assemblages. Where the landforms displayed clearer evidence of origin, a more genetic nomenclature was employed (e.g. lateral moraine). The topographic, sedimentological and morphological evidence discussed above is interpreted as follows: The sorted sediments,

which form a minor component of the deposits, are discontinuous, folded and faulted. These are interpreted as allocthonous materials. The presence of boulders embedded in, but protruding above the surface of, the deposits suggests that they were released from, or lowered by, a relatively inactive ice body. They were apparently not deposited in a subglacial environment.

PITTED MORAINE

This unit is the equivalent of kame-and-kettle topography described by Flint (1971, p.212). It is recognized that there are gradational forms between pitted moraines, hummocky moraines and kettled outwash, but pitted moraines are characterized by steep-sided, sharply outlined, closed depressions (kettles) together with sharply-defined positive relief features. The internal composition is dominated by till but stratified and sorted sediments comprise subordinate portions of these deposits. Pitted moraines are distinguished from hummocky moraines in that closed depressions characterize pitted moraines whereas rounded positive relief features are the distinguishing characteristics of the latter. In general,they differ from kettled outwash by displaying positive as well as negative relief features whereas pitted outwash trains are typically planar, containing only kettles. The sedimentary characteristics also differ between these three landforms. Pitted moraines are typically comprised of coarse-textured, poorly consolidated till or, in some cases, ice contact deposits, whereas hummocky moraines typically are comprised of a finer textured till matrix with little sorted or stratified material. Kettled outwash material is always stratified, although not always well sorted.

Pitted moraines are indicative of depositional environments where:

(1) supraglacial and englacial sediments have been draped over blocks of ice,

(2) blocks of ice have been partially or entirely covered with drift, or,

(3) thin ice has stagnated and irregularly downwasted adjacent to areas of active ice.

For the last case the nature of the ice remnant and the supraglacial debris must be such that uncontrolled disintegration, as described by Gravenor and Kupsch (1959), occurred thereby creating short, irregular ridges and hills as well as kettles.

Shallow ponds and muskegs are currently common in the kettle basins. These result from the large regional moisture surplus and the reduced infiltration rate of the till

compared to the coarser textured outwash deposits. The presence of organic terrain or ponds is therefore a useful indicator to differentiate between pitted moraines and pitted outwash plains. The presence of pitted moraines indicates that ice stagnation was dominant in the related areas during the final phase of deglaciation.

The largest concentration of pitted moraines in the study area is along the eastern slopes of the outer foothill ridges and within the easternmost foothills valleys. One such occurrence is found in the valley north of Brown Creek. The elevation of this pitted moraine is about 1520 m and it grades into a hummocky moraine at lower elevations. This suggests that the distributary lobe of ice which occupied the valley thinned and stagnated in these areas but lower in the valleys deposition may have been more controlled. The valley morphology and the meltwater channels within the valley indicate a flow toward the southeast. This suggests that the ice which occupied the valley was part of the Brazeau Lobe. Similarly, ice stagnation must have occurred in the area of the confluence of Crooked Creek and the Pembina River as a pitted moraine is found along the wall of the Pembina Valley at an elevation of about 1330 m. Again, the lower parts of the valley contain tracts of hummocky moraine indicating that uncontrolled disintegration was more prevalent along the valley sides, especially in ice-marginal or more distal areas.

Along the eastern slopes of the Foothills, north of the Pembina River valley, a pitted moraine occurs around the southeast side of Lovett Ridge. Here, the pitted moraine is near the valley bottom and it is adjacent to an ice-moulded moraine as well as hummocky moraine. However, the hummocky moraine extends to higher elevations on the eastern slopes of Lovett Ridge. A plausible scenario to explain this is that the hummocky moraine was dumped at the ice margin by a relatively active glacier which was flowing southeastward along the eastern margin of the Foothills. The origin and activity of the glacier are indicated by the ice-moulded moraine at lower elevations. As the ice-marginal debris thickened during wastage of the lobe,the debris mantle became instrumental in reducing the activity of the ice. The concentration of ice-marginal and supraglacial material along the interpreted ice margin appears to have been intensive as the subglacial morainic forms to the east do not display evidence of a mantle of supraglacially or englacially-derived material. This suggests that during ice wastage there was still significant local ice activity.

The easternmost pitted moraine lies immediately north of the Brazeau River valley and is associated with an esker complex to the east of the Foothills. According to local ice-flow indicators these moraines are on the down-ice side of the Tertiary uplands. It is probable that ice stagnation in this area involved the progressive isolation of thin ice down-glacier of the partial topographic barriers created by the uplands. It is also possible that mutual interference between ice flowing south from the Athabasca Valley and a lobe of ice extending from the Brazeau River valley resulted in a prolonged period of ice stagnation in this location.

Within the Foothills an area of pitted moraine occurs adjacent to the large esker complex near Thunder Lake (see Landforms Map). The extensive esker complex in this area indicates that ice stagnation occurred probably as the stresses in the Brazeau Lobe became incapable of driving the flow around the local, 45 degree angle, bend in the valley. No extensive glaciolacustrine sediments were found in this area, thus it is assumed that the pitted moraine resulted from the melt-out of ice blocks buried by supraglacial material in an environment where adequate drainage was available. If stagnation had occurred due to obstruction of the Brazeau River valley by Athabasca ice, one would expect to find evidence of ponding as well as ice stagnation. The adjacent esker complex (see Fig. 3.9b) and the pitted moraine indicate that although some ponding probably did take place slightly downvalley of this location, there was an uninterrupted flow of meltwater through this part of the valley.

Pitted moraine of a probably similar origin occurs at the narrowest part of the valley where the Brazeau River transects the Front Range. The valley in this area also bends through an angle of nearly 45 degrees, but in this case the ice stagnation features occur on the inside of the bend at the constriction in the valley. On the outside of the curve active ice landforms (drumlinoid ridges) occupy the valley floor. This landform assemblage suggests that the resistance to flow caused by the sharp bend produced ice stagnation along the inside of the bend. Supraglacial debris probably buried blocks of ice on the inside of the curve.

The conclusions which can be drawn from the pitted moraines mapped in this study are as follows:

(1) Uncontrolled ice disintegration, as described by Gravenor and Kupsch (1959),

played a subordinate role in the geomorphic evolution of the area. Only six pitted moraines were identified and five of these were located within the Foothills portion of the study area.

(2) The elevations of these landforms varies from about 1250 m to about 1550 metres. In most instances they grade into other morainic landforms which suggests that they were not all created by one synchronous regional stagnation.

(3) The common pattern of pitted moraine grading into hummocky moraine, or vice versa, indicates that progressive or incremental stagnation played a significant role during deglaciation. This is augmented in the case where ice-moulded moraine lies adjacent to pitted moraine. The lack of an obvious moraine mantle over the active-ice landforms indicates that the ice retained sufficient activity during the formation of the pitted moraine to concentrate material at the ice margin in order to produce the pitted moraine.

CREVASSE FILLINGS

Superimposed on some of the ice-moulded landforms that are found on the plains to the east of the Foothills, are relatively small (2 – 5 m high), straight ridge complexes. In some respects the morphology of these features resembles eskers but they differ in that the ridges tend to form a trellis pattern with straight ridge segments and ridge junction angles of approximately 90 degrees.

These features are relatively inconspicuous from the surface, or from low altitudes, because of the heavy forest cover. However, they may be distinguished from higher altitudes, or on aerial photographs, because of the enhanced drainage which results from the positive relief of the features (see Fig. 3.6). The improved drainage causes a vegetational change (usually more vigorous growth) which accentuates the appearance of the ridges. Associated with the major ridge complexes (which are depicted on the Landforms Map) are minor ones, too small to be mapped. Generally the major ridges are parallel or sub-parallel to the local ice-moulded landforms, with the intersecting minor ridges lying normal to the inferred direction of ice movement. The ridges tend to be draped or superimposed on the underlying glacial landforms. Figure 3.6 illustrates ice-moulded ridges in association with a complex of ridges which have been

interpreted as crevasse fillings. The reticulate pattern observed in Fig. 3.6 is similar to that illustrated by Mollard (1982) from central Saskatchewan, east-central Alberta, and the southwestern Northwest Territories.

The sediments of these ridge systems have not been adequately investigated because they generally occur in relatively inaccessible areas and no exposures revealing the sedimentary characteristics of these landforms were found. Shallow, hand-excavated pits revealed poorly sorted, unstratified gravel and sand but it is not known if this material is typical of these landforms. Similar landforms described by other researchers contain either till (McPherson and Gardner, 1969; Hazelton, 1978) or a complex combination of glacial and glaciofluvial sediments (Lawson, 1979; Mollard, 1982).

The superimposition of these ridges over other identifiable glacial landforms, sometimes other ridges of the same complex, has been noted by other authors (McPherson and Gardner, 1969; Hazelton, 1978; and Mollard, 1982). Studies of contemporary glaciers (McPherson and Gardner, 1969; Hartshorn and Ashley, 1972) suggest that these landforms are created by extrusion of unfrozen saturated sediments into subglacial crevasses. This explanation appears to be the most satisfactory for these features as well because there is no evidence to suggest that they were formed in any time-transgressive manner such as at the terminus of a back-wasting glacier. If this interpretation is correct, these ridges reflect the crevasse pattern of the ice subsequent to local stagnation, the orientation of the major ridges may therefore provide useful information regarding the stress concentrations within the ice during wastage.

The formation of supraglacial or subglacial crevasses is dependent upon a tensional stress in the ice which may be caused by either a lateral or a longitudinal extension of the glacier. In order for subglacial crevasses to persist for sufficient time to allow sediment to be injected into the cavities, the glacier must have thinned to the point where the compressive stress at the base of the glacier was insufficient to allow the glacier to close tensional cracks by plastic deformation.

This interpretation is compatible with the observed evidence that these features are commonly found superimposed on ice-moulded landforms. They therefore must have been formed subsequent to the event(s) which created the ice-moulded landforms.

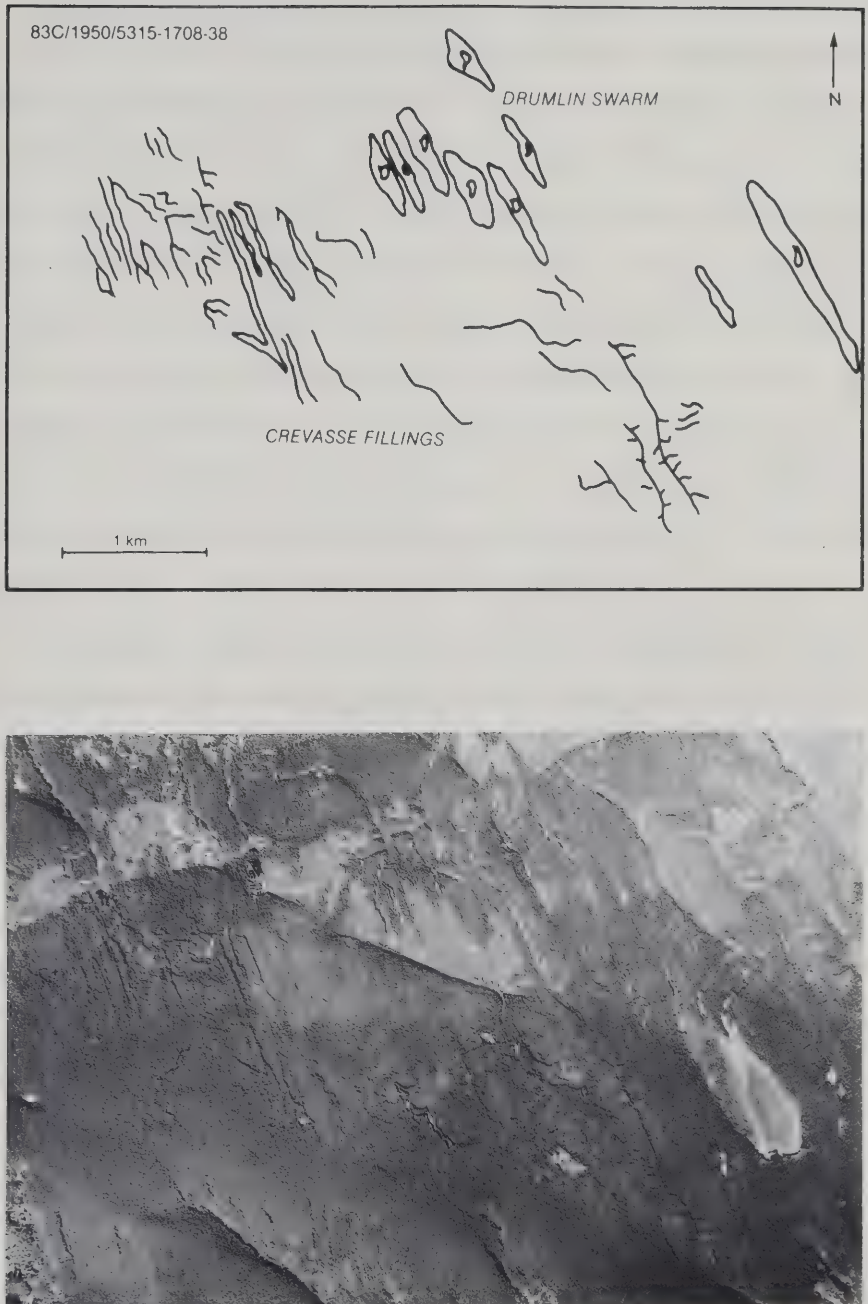


Fig. 3.6 Crevasse fillings in association with ice-moulded landforms.

As mentioned previously, the dominant ridge pattern in the Upland areas is parallel to that of the ice-moulded landforms. The ridge pattern reflects a dominant orientation sub-parallel to the interpreted direction of ice flow with a minor secondary component which is approximately normal to the flow direction. McPherson and Gardner (1969) described similar ridges which had recently been formed by the Saskatchewan Glacier and interpreted the linear ridges parallel to the valley axis as flutings and the transverse ridges as segments of annual moraines. Their interpretation does not seem useful in this study because the kinked crestlines of the ridge systems and the sharp-angle bends cast serious doubt on an interpretation which requires them to be formed by active ice.

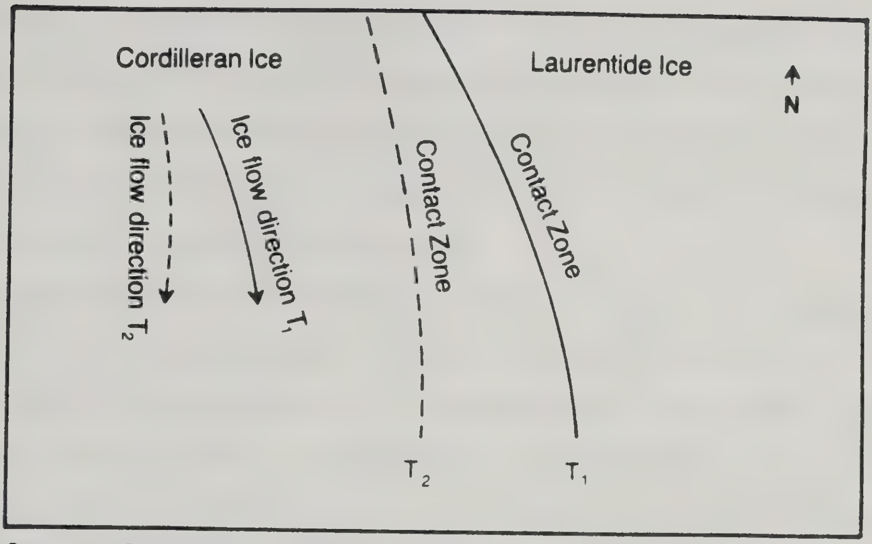
Gravenor and Kupsch (1959) provide more insight into possible genesis in their discussion of controlled versus uncontrolled ice disintegration. In their model controlled disintegration occurs when stress remains within the ice during the final stages of melting. The landform pattern created by controlled disintegration would then bear a direct relationship to the residual stress pattern within the ice and would tend to be oriented parallel to (splay crevasse pattern), transverse to (thrust-plane crevasse pattern), or 45 degrees to (chevron crevasse pattern) the direction of ice movement. In contrast, during uncontrolled disintegration, the glacier tends to break up with no apparent linear pattern and therefore produces a suite of oval, hexagonal or polygonal features.

If the model described by Gravenor and Kupsch (1959), Hartshorn and Ashley (1972), and illustrated by Mollard (1982) is accepted, then, on the basis of the sedimentary characteristics and the morphology of these ridges, it may be assumed that they were formed by the extrusion of subglacial sediments or the entrapment of glacigenic sediments within crevasses formed during the late stages of local ice disintegration.

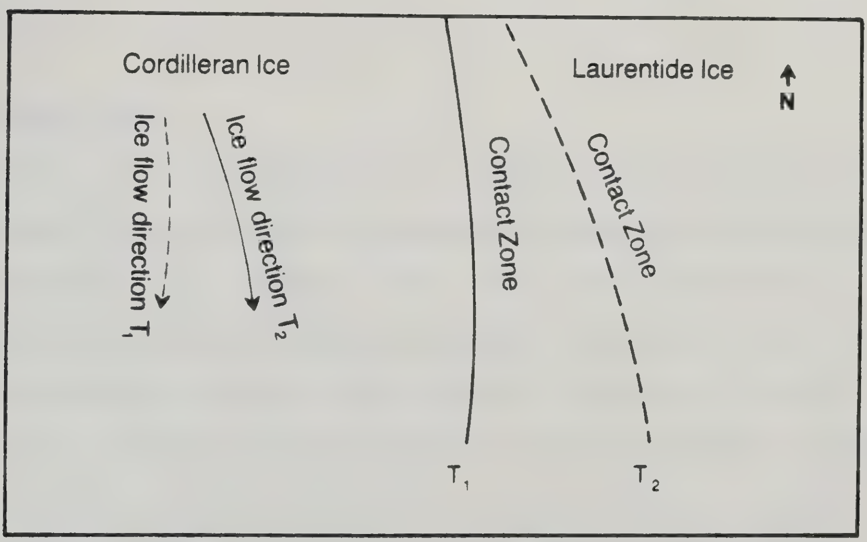
The fact that crevasses form as a result of brittle fracture of ice under tension (Nye, 1952) allows some interesting speculation concerning the regional pattern of deglaciation. A close examination of the orientation of the crevasse fillings and the ice-moulded landforms indicates that in the area depicted in Figure 3.6 the crevasse fillings are oriented 337 degrees to 157 degrees whereas the ice-moulded landforms are oriented about 332 degrees to 152 degrees. If the crevasses which formed the major crevasse fillings were parallel to the direction of flow immediately prior to

stagnation, then this flow was directed slightly more toward the Foothills than the earlier flow which produced the drumlins and flutings. This would occur only if the Laurentide ice to the east downwasted more slowly than the Cordilleran ice. If they downwasted simultaneously the lateral expansion of the Cordilleran ice, which is presumed to have produced the dominant crevasse pattern, would have been accordant with the ice-moulded landforms. However, if the Laurentide ice wastage preceded the disintegration of Cordilleran ice the crevasse pattern would have been oriented more toward the east as regional topography exerted a larger influence over the direction of ice movement. Figure 3.7 may help to illustrate this concept. It should be noted that the observed difference in orientation is only five degrees and there are other mechanisms which might account for this difference. Therefore, this interpretation must be considered to be tentative.

The crevasse fillings in the Upland areas (see Fig. 3.6) and those immediately east of the Foothills between the Brazeau and Pembina rivers appear to have been formed by a different crevasse pattern. The curved ridges between the Pembina and the Brazeau rivers (most westerly examples depicted on Landforms Map) could be explained in terms of a chevron crevasse pattern produced by the resistance to flow (drag) generated by the eastern margin of the Foothills on the southeastward flow of ice from the Athabasca Valley. In fact, a close examination of the Landforms Map reveals that they occur immediately downflow (southeast) of the foothills ridge where Fairfax Lake is situated. Following the models for crevasse propagation described by Nye (1952) and others, it is reasonable to expect crevassing at this location because two processes may have acted independently or collectively to generate crevasses. As mentioned, these crevasse fillings may have been produced by chevron crevasses but this is not likely as it is difficult to imagine chevron crevasses extending to the base of a glacier during deglaciation. The crevasses most likely reflect a lateral expansion of the Athabasca Lobe into the valley currently occupied by the Pembina River. Following the argument presented above, this would imply that the ice within the Foothills, except for the major valleys such as the Brazeau River valley, deglaciated earlier than the area immediately east of the Foothills. If this were not the case, and ice from the Pembina Valley merged with Athabasca ice, the mutual interference of the two ice bodies would be expected to result in laterally



Case 1: If Cordilleran ice wastage preceded Laurentide ice wastage, the direction of flow would deflect clockwise toward the Foothills.



Case 2: If Laurentide ice wastage preceded Cordilleran ice wastage, the direction of flow would deflect counter-clockwise away from the Foothills.

Fig. 3.7 Theoretical changes in the direction of flow of Cordilleran ice which would be induced by different sequences of deglaciation

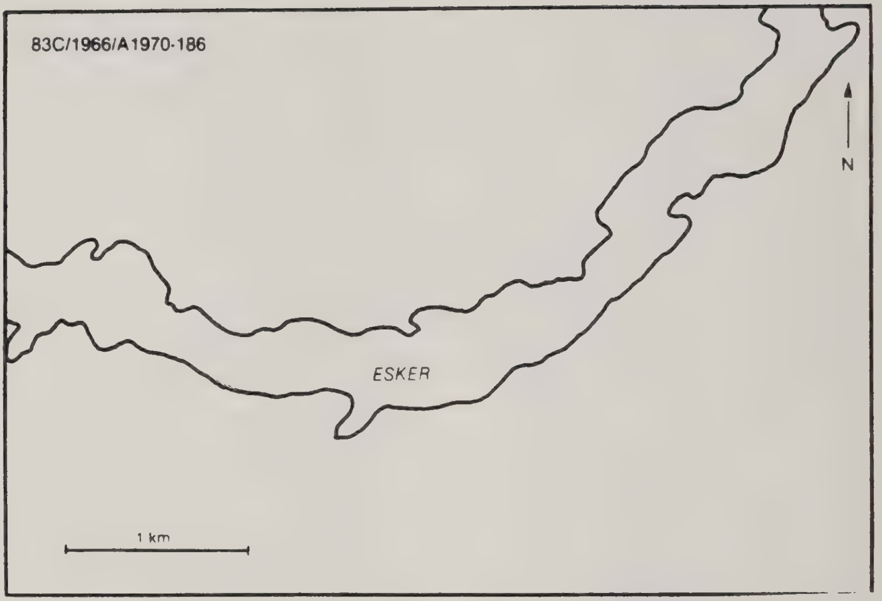
compressive flow of the Athabasca glacier which would preclude the formation of crevasses in this area. The inference expressed earlier, that the orientation of some of the landforms suggests that the Athabasca glacier downwasted before the Laurentide glacier, is supported by both the existence and the orientations of the crevasse fillings across the eastern portion of the study area.

The presence of several suites of crevasse fillings along the upland plateaus of the Western Plains indicates the relative importance of ice stagnation, as opposed to frontal retreat, during deglaciation. The preservation of these small ridge structures suggests that the sediments were deposited in crevasses that formed in inactive ice. The evidence of local divergence, such as the crevasse fillings near the eastern foothills margin, illustrates that there was local divergence within the valleys during deglaciation. The stagnation was therefore probably incremental stagnation rather than regional stagnation.

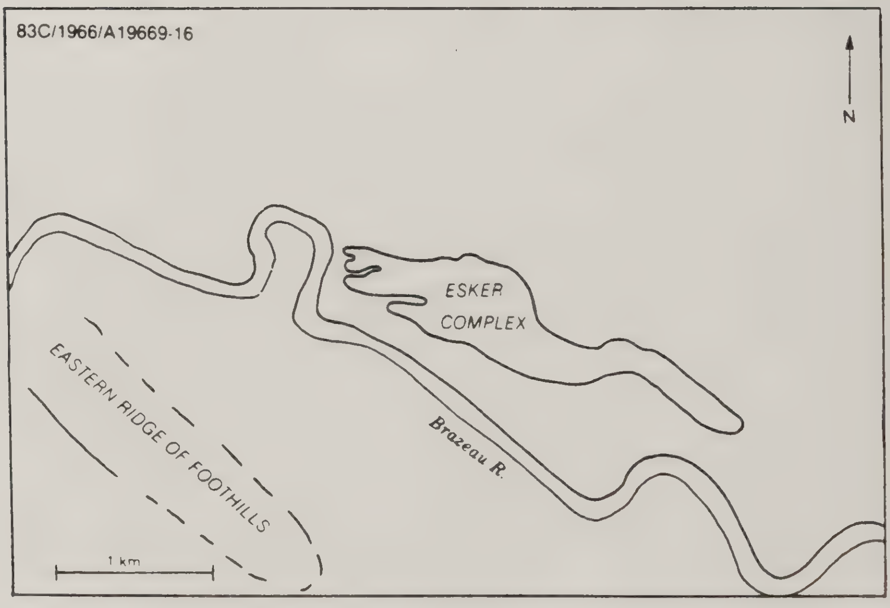
ESKER COMPLEXES

Eskers are rarely mentioned in the literature on the Foothills and Mountains of Alberta. Roed (1975, p.1512) briefly mentioned the Emerson Lakes esker complex which lies to the east of the Athabasca River, east of the Foothills. Rutter (1972, p. 23) describes eskers between Kananaskis and Seebe in the Bow River Valley. Gambier (1984) expanded on Roed's discussion and described the sedimentary structures found within the Emerson Lakes eskers. Boydell (1978) did not mention eskers in the Rocky Mountain House area but his map legend includes a unit of 'glaciofluvial deposits', which may include eskers. McPherson (1970) described the glacial deposits within part of the Front Ranges and he did not identify eskers in the upper North Saskatchewan River Valley. Therefore, it is noteworthy that in the present study five large esker complexes were identified, all of which are more than 3 km long. In addition, one other esker was found that was too small to be mapped at a scale of 1:100,000 (Landforms map scale).

The composition, topographic position, and morphology of the eskers vary considerably. Two eskers are located to the east of the Foothills. One occupies the flat lowland area between the Blackstone and Brazeau Rivers (Fig. 3.8a) and the other lies along the north rim of the Brazeau Valley in the hilly uplands (Fig. 3.8b). The lowland esker



83C/1966/A1970-186
ESKER
N
1 km



83C/1966/A19669-16
N
ESKER
COMPLEX
EASTERN RIDGE OF FOOTHILLS
Brazeau R.
1 km

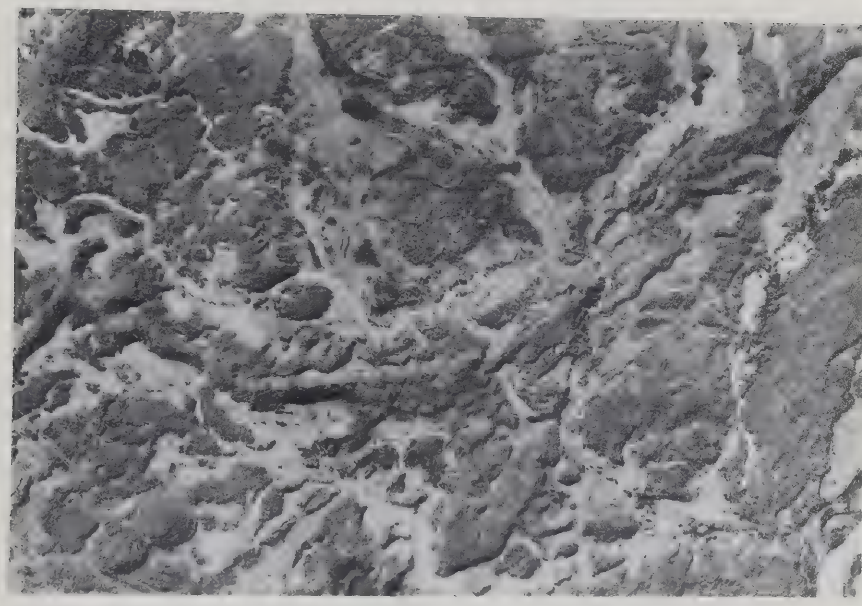


Fig. 3.8(a) Esker on the Western Alberta Plains, between the Blackstone and Brazeau rivers.

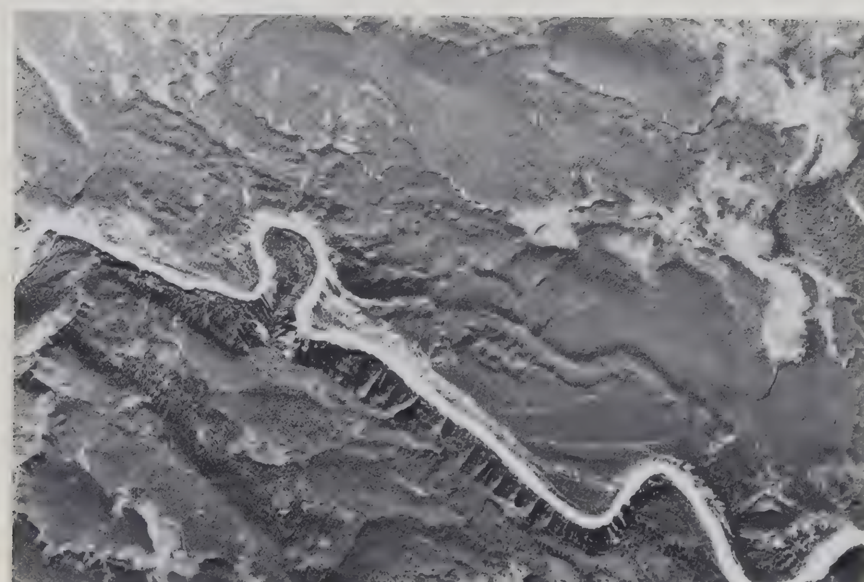
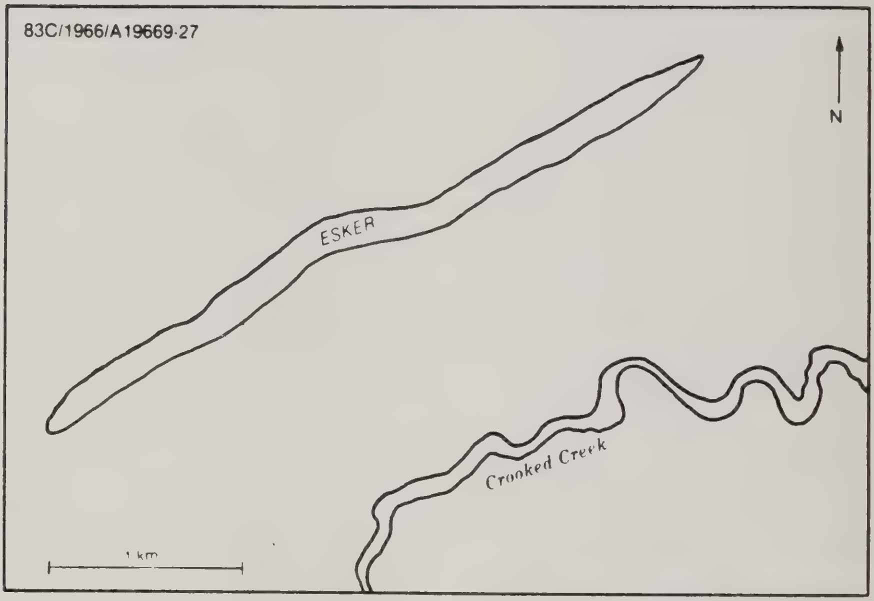


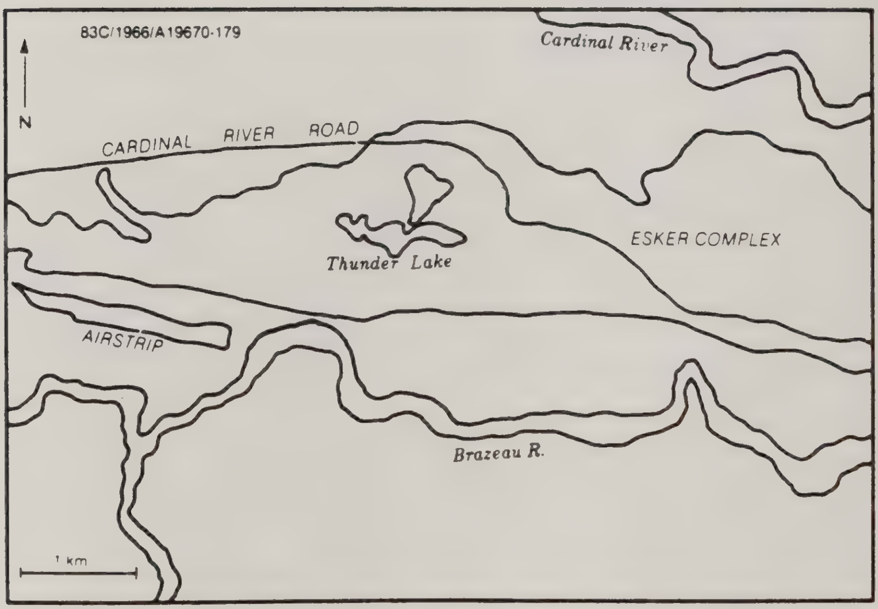
Fig. 3.8(b) Esker on north bank of the Brazeau River East of the Foothills.

(Fig. 3.8a) is an irregularly–shaped ridge with moderate (10 –15 degree) slopes and irregularly–shaped kettles along the flanks of the ridge. The material comprising the ridge is predominantly medium–textured sand with some fine gravel and occasional cobbles. The upland esker (Fig. 3.8b) consists of a number of anastomosing ridges and pitted plateau surfaces which coalesce to form a single, well–defined ridge of coarse gravel. Slope angles are moderately steep (25 – 28 degrees) along the major ridges. Sorting is variable with sands and gravel lenses observed in exposures of the anastomosing section and coarse (7–12 cm in diameter) gravel comprising the material of the major ridge structures.

Two large eskers lie on either end of the Cardinal Hills. The northern one is a steep–sided (up to 31 degrees) ridge, approximately 20 m high, which parallels the Pembina River and lies adjacent to Crooked Creek, a tributary to the Pembina River (Fig. 3.9a). Few exposures exist so that the constituent material is largely unknown, but well–rounded clasts greater than 50 cm in diameter were observed in shallow exposures created by wind–thrown trees. A borrow pit at the western end of the esker revealed medium–textured sand with rounded clasts up to 5 cm in diameter. The surface of this esker is mantled by a thin veneer of diamicton, probably of ablation moraine origin.

The esker to the south of the Cardinal Hills occupies part of the area between the Cardinal and Brazeau Rivers, upstream from their confluence. Thunder Lake lies within this esker complex. The Thunder Lake esker (Fig. 3.9b) is the largest esker complex so far documented in Alberta. It can be traced for more than 14.5 km along the Brazeau River valley and in places it is more than 2 km wide. The surface morphology and sedimentary characteristics of this esker are variable but, in general, it consists of steep–sided ridges (up to 31 degrees) of coarse gravel with a sand and fine gravel matrix. Relative relief is commonly as great as 40 m and generally the ridges parallel the flow direction of the present Brazeau River. Anastomosing ridges are frequent, with the most discrete and clearly defined ridges adjacent to the present river valley and the more chaotic ridge pattern, with an increased incidence of kettles, along the north side of the valley near the Cardinal River. The texture of the deposits also changes from coarse material, in which rounded clasts up to 30 cm in diameter are found, to coarse sand and fine gravel in the pitted area along the side of the valley. The lithologies identified in this material are


83C/1966/A19669-27
N
ESKER
Crooked Creek
1 km


83C/1966/A19670-179
N
Cardinal River
CARDINAL RIVER ROAD
ESKER COMPLEX
Thunder Lake
AIRSTRIP
Brazeau R.
1 km

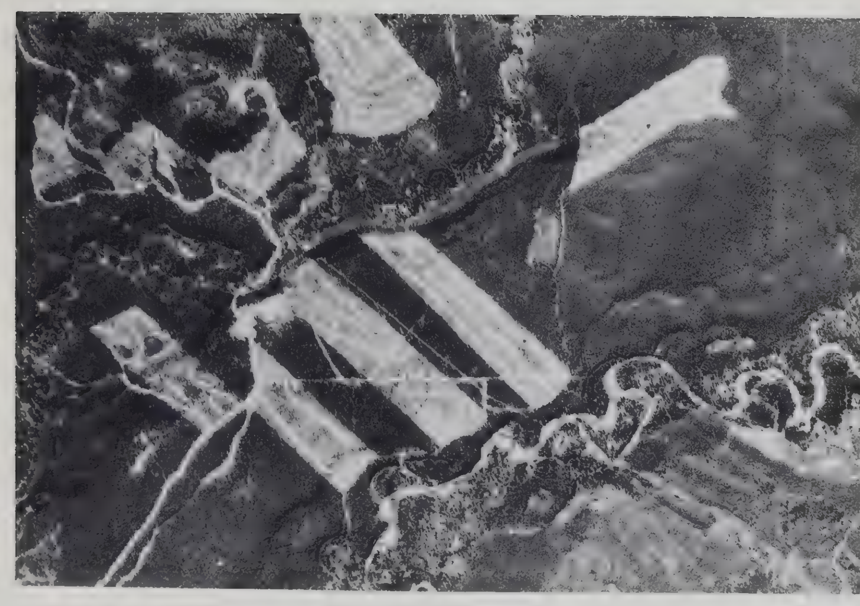


Fig. 3.9(a) Esker adjacent to Crooked Creek along the south side of the Pembina River Valley within the Foothills.

Fig. 3.9(b) Thunder Lake esker complex.

predominantly limestones and quartzites but conglomerates, sandstones and siltstones are also present. A thin, discontinuous mantle of diamicton overlies the sand and gravel deposits along the northern and eastern portions of the complex. It is probably this fine textured material, together with the underlying shale, which has decreased the permeability of these deposits sufficiently to permit ponding in some of the larger kettles.

The fifth large esker complex occurs immediately east of the First Range in the structural valley between the mountain front and the Foothills (Fig. 3.10). It extends northward from the Southesk Valley and begins as a linear ridge complex, with well-defined kettles along the flanks of the 10 m high ridges. It trends northward into a chaotic complex of ridges and kettles. The ridge configuration indicates that the related meltwater flow was bifurcated by local topography as the western ridges extend north along the Neilson Creek valley and the eastern ridges parallel the present Southesk Valley. Again, a thin veneer of fine-textured diamicton mantles the surface of most of this esker. In most exposures this mantle was less than one metre thick, the largest clasts were less than 30 cm in diameter and usually sub-rounded in shape. The matrix consisted of silt and fine sand. The internal composition of the esker appears to be almost entirely medium and coarse textured gravel (up to 80 cm in diameter, with a modal diameter of 10 cm). Clasts tend to be sub-rounded to sub-angular with limestone and quartzite being the dominant lithologies. A medium-textured sand matrix is characteristic and in the upper 50 cm a heavy carbonate precipitate is cemented to some of the clasts. The observed differences in morphology of these features is potentially significant. It has been suggested (Rutter pers. comm. 1981) that the slope angles, which vary from 10 to 15 degrees in the subdued lowland esker (Fig. 3.8a) to as steep as 31 degrees within the Foothills (Fig. 3.9a) reflect eskers of different ages. It is possible that these differences are an indication that the lowland eskers and the upland esker east of the Foothills are older than the eskers further to the west. Alternatively, the slope differences may reflect differences in the depositional environment at the time of formation and the ages of the eskers are not significantly different. To resolve this problem, other lines of evidence must be reviewed because slope angles in coarse-textured sediments are not particularly useful to demonstrate differences in age between eskers because mass movement

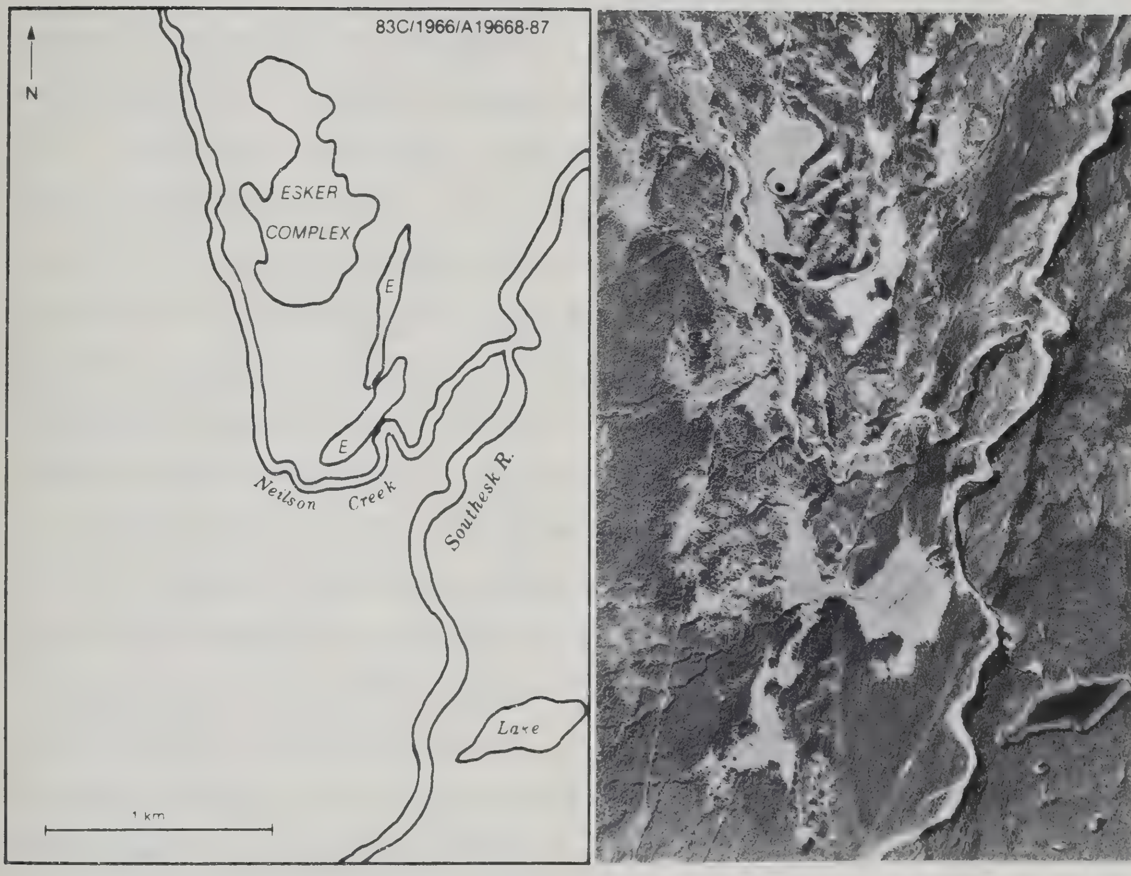


Fig. 3.10 Esker Complex between the Southesk River and Neilson Creek, western margin of the Foothills.

processes tend to be retarded in coarse-textured, well-drained materials. The very fact that eskers survive as steep-sided ridges after having been subjected to an unknown, but complex, set of processes associated with deglaciation and the early post-glacial interval suggests that deglaciation must have been a relatively rapid and continuous process.

If the surfaces are of different ages, in addition to subdued slope angles, differences in staining and weathering of surface clasts, soil profile development, loess thickness and carbonate leaching should be observed. Field observations indicate no obvious differences in these parameters. In fact, the greatest amount of carbonate translocation was observed in the Southesk esker complex (Fig. 3.10) which is adjacent to the mountain front.

The existence of esker complexes, their configuration, and locations are highly relevant to the geomorphic interpretation of the area. Eskers which are mantled with till, and not beaded, are indicative of deposition beneath inactive ice (Embleton and King, 1975, p.473). This interpretation is appealing but perhaps too simplistic. The large Thunder Lake esker and, to some extent, the Neilson Creek esker and Blackstone River esker (Fig. 3.8a), have ridge systems of coarser-textured material but they grade laterally or downvalley into finer-textured, kettled deposits. It is not obvious how englacially or sub-glacially channelled meltwater could decrease in velocity sufficiently to deposit fine sand and silt, especially in light of Shreve's (1972) study which illustrates that secondary currents in glacial conduits tend to move material inward and upward toward the center of the tunnel. Other researcher's (Banerjee and McDonald, 1975; Ringrose, 1982 and Saunderson, 1982) envision that the sedimentary evidence indicates that most, if not all, of the esker sediment was transported during times when the conduit was completely filled with water and sediment. This would explain the variability in texture observed within this study as the material would be moving as a slurry and not as bedload or suspended sediment as we traditionally have supposed. This is similar to the sliding bed facies described by Saunderson (1982). It would also explain some of the extremely coarse textures (clasts commonly in the 60 - 80 cm range) observed in the esker ridges but it offers little insight into the problem of why these features grade into more planar landforms comprised of finer-textured deposits. A further complication is the thin diamicton mantle which overlies at least some of these sediments. The diamictic unit suggests that either these landforms were constructed entirely subglacially or that flows of debris were transported onto these surfaces by a subsequent event.

The interpretive implications of this dilemma are best expressed in the Thunder Lake esker example. The ridges of coarse-textured facies grade downvalley into finer sediments with a flattened upper surface which is marked by scattered kettles. Morphologically it appears that a channelled flow (represented by the esker ridges) discharged into a delta-like environment which contained blocks of ice. Ringrose's (1982) model would explain the ridges as areas where a steeper hydraulic gradient and more rapid and perhaps episodic discharge of meltwater production existed. The laterally developed esker segments comprised of finer sediments would have formed under

conditions of lower flow and more uniform velocities (shallower hydraulic gradient) where the channels would be able to erode laterally as the channel bed aggraded; a situation similar to a subaerial braided stream.

Further down-valley, at approximately an accordant elevation, are glaciolacustrine sediments (see section B9/R, Chapter 4) which would be consistent with the interpretation of these sediments representing a delta-like environment. Such complex patterns of deposits are not unprecedented (Carruthers, 1953; John, 1972) but a satisfactory interpretation remains somewhat enigmatic. A subglacial delta does not seem plausible given Shreve's (1972) study which illustrates that channelled flow will be the norm. A re-advance of the glacier in order to deposit a thin mantle of till would certainly destroy the ridge structures of the eskers. It would therefore seem most likely that the esker ridge systems formed in a subglacial environment. The distal portion of the Thunder Lake esker probably represents a proglacial or ice-marginal delta which grades downvalley into a lacustrine environment. This explanation requires meltwater ponding within the Brazeau River valley adjacent to the Cardinal Hills; a requirement which is quite compatible with the other landform evidence mapped.

In any case, it is quite clear that the eskers are late-glacial features which signify at least local ice stagnation. Flint's (1971) conclusion that large eskers tend to form in subglacial passages when meltwater discharge is high and the hydraulic gradient is decreasing seems to have passed the test of time. Shreve (1972) describes how eskers may form beneath active ice but it is difficult to believe that they could be preserved in such an environment. It is therefore much more likely that the eskers observed on the landscape were exposed from beneath relatively inactive ice.

The distribution of eskers is evidence that ice stagnation occurred to the east of the Foothills along the western side of the Athabasca Lobe, in both the Pembina and Brazeau River valley and against the bedrock ridge close to the exit of the Southesk River from the mountains. Parenthetically, it should be stated that an esker too small to be mapped (200 m X 30 m) has been mined for gravel along the Forestry Trunk Road about 1 km southeast of the Brazeau River bridge. It is oriented northeast-southwest and is at an elevation of 1335 metres. This is approximately accordant with the Thunder Lake esker complex up-valley, which suggests that the Brazeau Lobe extended to at least the

eastern side of the Foothills at the time of ice stagnation.

The direction of meltwater flow at the time of esker deposition is generally consistent with present drainage, as might be expected. One exception is the bifurcated flow indicated by the esker near the mountain front. The flow here appears to have been northward through Neilson Creek, opposite to the contemporary flow of that creek. Another exception is the lowland esker between the Blackstone and Brazeau Rivers. It appears to have been deposited by a stream that breached the Foothills ridge on the east side of the Forestry Trunk Road valley and flowed northeastward toward the present Brazeau River valley.

3.4.2 GLACIOFLUVIAL LANDFORMS

Four categories of glaciofluvial landforms are included on the Landforms Map. Meltwater channels are dominantly erosional landforms, outwash terraces are both aggradational and degradational features, and outwash plains and pitted outwash are depositional landforms.

MELTWATER CHANNELS

The geologic structure of the Foothills is oriented transversely to the general direction of interpreted flow of the Brazeau Valley glacier, as indicated by the landform assemblages within the study area. Similarly, the upland areas east of the Foothills generally are oriented transversely to the interpreted flow direction of the Athabasca Lobe. Because of this, and the intermediate physical relief of the area, discontinuous meltwater channels are extremely common. Classic examples of both ice-marginal and proglacial channels were found but, in order to restrict the number of landform units on the map, these have been combined into a single category. The presence of eskers in both the Foothills and the Western Alberta Plains portion of the study area provides indirect evidence that considerable amounts of englacial or subglacial meltwater flow occurred. Rothlisberger's (1972) and Shreve's (1972) studies indicate that some subglacial channels (Nye channels) probably formed but if the physical form of these channels survived deglaciation, they escaped recognition during this study.

Because of the abundant moisture availability of the contemporary environment, most of the meltwater channels presently encompass bog areas, sloughs and shallow lakes or ephemeral streams. Thus they are easily identified from airphotos or by aerial reconnaissance. Typically, the meltwater channels are not well-developed channel systems but rather are discontinuous channel segments. Only the more obvious examples of meltwater channels were mapped in this study. Even these channels are less deeply incised and more segmented than the meltwater channel systems of the areas to the east which were formed by retreating Laurentide ice (e.g. the Battle River, Vermilion River, and Grizzly Bear Creek systems). Perhaps this reflect the differences in ice thickness and therefore the volume of meltwater discharge produced in the two areas. Glaciers tend to achieve equilibrium when their upper surfaces are accordant. The Western Alberta Plains are approximately two hundred metres higher in elevation than the area immediately east of the North Saskatchewan River at Drayton Valley, which is directly to the east of the study area. Ignoring isostatic adjustments for ice load, this would represent a considerable difference in ice thickness and, by implication, the potential amount of meltwater produced during deglaciation. Another possible explanation is that foehn winds may have increased the amount of ice-wastage which occurred due to sublimation in the area immediately to the east of the mountain front. A problem with this argument is that the assumed surface profile of the Laurentide glacier would have counteracted the foehn effect further to the east (see Rains and Kvill, in press, for assumed surface profiles during Late Wisconsin maximum).

The somewhat deranged and discontinuous nature of the meltwater drainage systems relative to the presumed large volume of ice, is rather puzzling. With the ablation of the extensive ice lobes, evidenced by the landforms discussed earlier, one might expect that major channels, at least as large as the Battle River valley which lies further east on the plains, would have been produced along the eastern edge of the Foothills and along the contact zone between Cordilleran and Laurentide ice. The only clear examples of integrated channel systems of this nature occur 80-100 km east-southeast of the study area, where now the underfit Medicine, Blindman and Battle Rivers occupy valleys which obviously carried large amounts of meltwater southeastward.

Even if one accepts the arguments that there was less meltwater produced in the area which was overridden by Cordilleran ice, the meltwater which was produced should influence the density of the drainage pattern and the size of the channels but it does not explain the poor channel morphology and the discontinuous nature of the channel systems. Perhaps deglaciation was sufficiently rapid that steep-sided channel systems formed but were obliterated by post-glacial processes such as mass movement. Perhaps significant amounts of meltwater were transported englacially or supraglacially and channel segments represent only the areas where the stream channels were in direct contact with the glacier bed. This obviously occurred in locations where meltwater channels notched structural ridges but it was less obviously the case to the east of the Foothills where the channel segments frequently occur in depressional sites and are not particularly common across the upland areas.

The drainage patterns illustrated by the meltwater channels are useful in reconstructing the pattern of deglaciation. To the east of the Cardinal Hills, where the Brazeau and Pembina Rivers almost converge, there is a series of meltwater channels oriented approximately northeast-southwest (see Landforms Map). These show that meltwater flowed from one basin to the other at some time during deglaciation. The topography and, in some places present drainage, suggest a flow from the west limb of the Brazeau River into the east limb of the Pembina River, but slope measurements taken within the relict channels show the opposite. Cross-bedded sediments measured in one of these channels (see section B16/L, Chapter 4) indicate a paleocurrent direction toward 200 degrees True (S.W.).

Meltwater appears to have been ponded between the eastern slopes of the Foothills and the upland areas on the plains to the east. Before the lowland to the northeast, presently drained by the east limb of the Pembina River, became ice-free, meltwater apparently breached the lower divide along the western end of the uplands between the two basins and exited via the Brazeau River system. A second line of evidence to support this hypothesis is that the Brazeau River valley, downstream from the meltwater channels, occupies a relatively broad valley with numerous terraces along the valley walls. Upstream of the relict meltwater channels the Brazeau River valley is essentially a 100 m deep gorge with steep, unstable walls.

If the hypothesis stated above is correct, the lower portion of the valley should have carried runoff from both the Pembina and the Brazeau basins (as well as meltwater from the Athabasca Lobe) whereas the Brazeau River, within the Foothills, carried only local drainage. As stated previously, east of the Foothills the relict meltwater channels tend to maintain a northwest-southeast orientation except for south of the Blackstone River. In this area the higher terrain of the Foothills to the southwest appears to have been the source area for some local streams which flowed northeastward into the Brazeau River system. Along the eastern flanks of the Foothills several ice-marginal channels are incised into bedrock. The sides of these channels are steep and canyon-like.

Within the foothills the channels radiate outward from the main axis of Brazeau River valley. For example, abandoned channel segments indicate that meltwater flowed from the Brazeau Valley into the Crooked Creek Valley and also into the Forestry Trunk Road Valley. This suggests that active ice must have occupied the main valley when the ice surfaces of adjacent areas were topographically lower than the ice surface along the main valley. This implies that ablation had lowered the surfaces of the less-active ice in the adjacent Foothills while the ice surface remained high along the main valley because it was being supplied from the mountain areas.

All channels observed seem to be a reflection of a general northwest to southeast hydraulic gradient which is consistent with the regional ice-flow indicators. The trend of the channel segments to the east of the Foothills also reflects an eastward component which indicates that the contact zone between Cordilleran and Laurentide ice was an important meltwater outlet. This is supported by the extensive fluvial deposits which may be observed in many of the roadcuts in the vicinity of the Brazeau Dam. This interpretation is consistent with the landform evidence and the ice profile models proposed by Rains and Kvill (in press).

An abandoned meltwater channel system supporting the interpretation offered above transects the ridges between the Moosehound Creek and Blanchard Creek drainage basins (see Landforms Map). From air photos and aerial investigations it appears that this is part of a stream system which carried a flow of water from the west side of what is now the Blanchard Creek basin to the confluence of the Brazeau and Forestry Trunk Road valleys. The valley train deposits which occupy the floor of the channel segments indicate

that the stream was depositing sediment and graded to an elevation of approximately 1380 m which is about 20 m below the lateral moraine that occupies the valley wall near the outlet of this stream channel. The flow direction of this valley indicates an ice gradient from west to east. Evidently this was an ice-marginal meltwater channel which was probably only active when supraglacial melting exceeded the discharge capacity of the englacial and subglacial systems.

In summary, there appears to be a much less integrated relict meltwater system within the study area than exists in the adjacent area to the east. Some of this is probably due to the difference in the amount of ice and local climatic conditions, but there is at least some evidence to suggest that the rate of deglaciation, or rapid changes in the ablation rate, prevented the development of an integrated channel system in this area. Alternatively, the rapid rates of recent erosion and mass movement may have destroyed much of the evidence of relict channel systems.

In any case, the meltwater channel patterns which can be traced reflect a general southeastward transport of meltwater with a transverse component which emerges from the Foothills in an easterly direction. Many of the channel segments within the Foothills document the fact that the main Brazeau River valley was filled with ice during the time when they were occupied by meltwater streams.

PITTED OUTWASH TRAINS

These are outwash deposits which have been deposited on top of, or around , inclusions of glacial ice. Subsequent melt-out of the ice resulted in an uneven and pitted surface. This unit is identified here as pitted outwash train instead of kame-and-kettle topography on the basis of the density of pits or kettles on the surface. The surface of these units is usually much less chaotic than that described and depicted by Embleton and King, (1975, p. 525) or Sugden and John (1976, p. 333).

Three units of pitted outwash were mapped. East of the Foothills is an extensive area of pitted outwash lying between the Brazeau and Blackstone Rivers. Sand and fine gravel are the major constituents of this unit. The surface topography is undulating and the kettles are large and irregularly shaped. This suggests that the glaciofluvial deposits were superimposed over a thin layer of stagnant ice similar to the situation observed on

the contemporary Saskatchewan glacier. A large esker system which lies along the west margin of this unit reinforces this interpretation.

Near the confluence of the Cardinal and Brazeau Rivers is a section of pitted outwash train that is again adjacent to an esker complex. The outwash unit would fulfill the text-book requirements of kame-and-kettle topography in that the surface is extremely chaotic, consisting of mounds and deep kettles. The material is coarse textured. Clasts in the 15 - 20 cm range comprise a relatively large per centage of the material although the modal size is closer to 2 - 2.5 centimetres. Generally, sorting is poor to moderate and slump and fault structures were observed in some sections. A mantle of matrix-dominated diamicton overlies most of this unit. Modal clast size is in the range of 4 cm but frequently they range as large as 40 centimetres. Sandstone and quartzite were the most common lithologies and many fractured clasts were observed in the diamicton mantle. The coarse texture, poor sorting and diamicton mantle suggest that the glaciofluvial deposits were laid on and around stagnant ice blocks but a glacier, or glacier remnant, was sufficiently close that debris could avalanche onto the surface.

Along Ruby Creek, immediately upstream of its confluence with the Cardinal River, a unit of pitted outwash occupies the bottom of the valley. The surface of this unit consists of irregularly-shaped ridges and small kettles. Sorting and stratification of the sediments are poor and the texture is in the fine to intermediate gravel range. The modal clast size is 1.5 - 2 centimetres with a very small per centage of the clasts exceeding 5 cm in diameter. Again, a diamicton mantles much of the surface and inclusions of unstratified sediments are common within the sorted deposits. Many examples of secondary structures such as faults and folds were observed in these deposits. This combination of evidence indicates that local blocks of stagnant ice must have been prevalent in this valley during deposition of the material.

OUTWASH TRAINS AND PLAINS

Much of the outwash in the study area has been incised by stream erosion to form terraces. Two areas of undissected outwash were mapped and these merit some discussion. The first occupies the floor of the Hanson Creek Valley. No good exposures were found along Hanson Creek but it appears that this outwash train is generally less

than 2 m thick and it directly overlies bedrock. The slope of the Hanson Creek Valley indicates a flow of meltwater away from the main valley, in the same manner as indicated by the meltwater channels discussed earlier. The difference is, of course, that in this valley meltwater was depositing sediment whereas in the channels it was eroding material.

The second outwash plain is located to the east of the Foothills and south of the Brazeau River. The sand and gravel which make up this unit are similar to those found in the kettled outwash to the north but morphologically this unit differs in that the surface is free of kettles. The southern end is mantled by, or grades into, lacustrine sediments. No stream exposures or man-made cuts display a complete section of this unit so it was not possible to determine the depth of these deposits. Evidently this unit represents deposits laid down in a glaciofluvial environment during a late stage of deglaciation and the lack of kettles or hummocks indicates that ice was not buried in, or beneath, the outwash.

OUTWASH TERRACES

Well-defined terraces flank the valleys of the Brazeau, Southesk and Cardinal River systems, as well as the lower reaches of many of their tributaries. The upper surfaces of these terraces were not surveyed so it is not possible to link them stratigraphically to glacial features up or down valley. They are identified as outwash terraces on the basis that terrace formation implies a high sediment load which, in this area, would imply proximity to a glacier. This interpretation is augmented by the coarse texture of the alluvium observed in exposures. Clast lithologies are primarily quartzite and sandstone but carbonates were also observed. Evidence of scour-and-fill structures was noted. Generally the sediments are moderately to well stratified but only moderately well sorted. A general fining upward was observed in some exposures but the texture is so coarse (clasts frequently as large as 60 cm in diameter) that no attempt was made to quantify this trend. Most clast shapes were sub-angular to sub-rounded.

In all three of the river systems mentioned above, one major terrace exists 50 – 60 m above the present river level. In all cases this appears to have been an extensive surface. Small channel scars on the surface indicate that braided channels dominated at the time of deposition. This probably reflects the abundant sediment supply and steep gradient which would have been present during late-glacial and early post-glacial time

when the surface was likely not stabilized by vegetation. This evidence, together with the fact that several abandoned meltwater channels grade into these terraces, suggests that deposition probably occurred during the final ablation of local ice. Upward fining of the sediments would be expected if a valley glacier was retreating upvalley removing the major sediment source from the area and reducing the amplitude of the fluctuations in discharge. McPherson (1970) and Boydell (1972) have described a similar upper terrace within the upper North Saskatchewan drainage basin.

Below this upper terrace level is a discontinuous, well-developed, paired terrace surface about 35 m above present river level. This is observable in the Brazeau and the Southesk River valleys but is lower (about 25 – 30 m above river level) in the Cardinal River valley. The surface morphology includes remnants of bars and channel scars which indicate that this surface was formed by a braided stream system. In addition to the two main terrace suites, there are numerous non-paired terraces in the valleys of these rivers, particularly along reaches that are not cut through bedrock ridges.

Stene (1966) mapped and interpreted two main terraces along the Athabasca River near Hinton. He correlated the formation of the upper terrace with a recessional moraine and the lower terrace to an unidentified position of the ice front further up valley. It is possible that his model is applicable to the terraces mapped in the present study but the landform evidence does not support this hypothesis. There is substantial geomorphic evidence of ice stagnation in this part part of the study area. Aggradation of outwash would be expected during an advance or still-stand of an active ice front, or distally from an area of ablating dead ice. No geomorphic evidence was found in the Brazeau River valley to support the argument for a still-stand position of an active ice front upvalley of the terraces. No end moraines were found to transect the lower parts of the major valleys. This renders the probability of these river terraces having been produced by aggradation associated with still-stand positions of a retreating ice front quite low. Those found near the valley heads were interpreted as being of being of late glacial or Holocene ages.

Glover (1979) studied the terraces along the Athabasca Valley between Hinton and Whitecourt. He argued that terraces may be produced by a combination of upstream and downstream influences and attributed at least three of the four paired terrace levels

identified to changes in the base-level of the Athabasca River brought about by the development of successively lower glacial lakes as the Laurentide ice retreated. Stene (1966) seems to have ignored the downstream influences which could have been partly responsible for the events which produced the terraces. The upper terrace surfaces of the Brazeau and Southesk Rivers, near their confluence, have relatively steep gradients. Elevations obtained from contours on the 1:50,000 scale NTS mapsheets indicate a gradient of approximately 0.006. At the confluence these terraces are approximately 183 m above the meltwater channels to the east of the Foothills and about the same height above the meltwater channels in the Forestry Trunk Road Valley. The upper terrace of the Cardinal River valley is about 155 m above these channels. It seems plausible that the terraces were formed by a sediment-laden river system flowing off dead ice in the tributary valleys of the foothills. Initially, an ice obstruction may have restricted the flow to the east of the Foothills, hence the development of a meltwater channel down the Forestry Trunk Road valley, but later the flow was carried eastward when the ice surface was lowered below the level of the notch eroded through the outer Foothills ridge (see Landforms Map). The second terrace may have been created by a period of incision, following a breach in the ice obstruction, and subsequent aggradation controlled by a similar topographic barrier or ice obstruction further to the east. Further research would be required to adequately support this hypothesis, but the fact that this terrace does not extend downstream of the Thunder Lake esker complex tends to augment the speculation that downstream controls may explain its formation. Terraces are found low in the Brazeau River valley, well to the east of the Foothills, but their topographic position makes it unlikely that they are continuations of those discussed previously.

3.4.3 GLACIOLACUSTRINE LANDFORMS:

GLACIAL LAKE BASINS

Obviously, with a large amount of ice ablating in an area of intermediate physical relief there is is likely to be considerable ponding of the meltwater. The highly variable nature of the interaction between ice and physical relief would have resulted in frequent changes in the location, size and shape of these ponds or lakes. Because of this, glacial lake margins are difficult to identify and there is a considerable degree of subjectivity in

interpreting the limits of these features. The policy adopted for this study was to map only those areas where lacustrine deposits produced an obvious alteration of the landscape. If the lacustrine deposits imposed a dominant influence on the present landscape the area was mapped as a glacial lake basin. If the lacustrine sediments formed only a thin veneer over the pre-existing relief, and the underlying landforms could still be identified, the unit was mapped with a compound designator such as lacustrine-mantled ground moraine. Many of the Alberta Research Council Surficial Geology Maps use a similar complex designator which uses color and patterns to denote an aeolian or lacustrine mantle superimposed on an identifiable substrate. It must be recognized that this procedure results in a conservative expression of the actual extent of lake sediments. Additional small areas of thin lacustrine deposits were noted in the field, especially along the eastern slopes of the Foothills, but these were not mapped. An over-reliance on airphoto mapping of lacustrine deposits in the Foothills should be avoided because valleys such as Grave Flats morphologically resemble glacial lake basins but in fact were not. The thin overburden in such valleys results in poor drainage which could be photo-misinterpreted as evidence of a glaciolacustrine deposit. Surface verification of lacustrine sediments is required for accurate classification and this is frequently complicated by the thick organic deposits which are prevalent in these areas. The few exposures which do exist indicate that silts and fine sand are the predominant textures but dropstones are very common. No rhythmites were observed and no beach deposits were identified. The sedimentary characteristics of these deposits are very complex. Beds of well-sorted material may contain facies of much coarser and less-well sorted material. Laminae were observed in some sections (e.g. B11/L, Chapter 4) and frequently the lacustrine unit is the uppermost unit in an exposed section of upward-fining sediments. Commonly significant changes in texture or sorting were observed without any obvious hiatus in sedimentation. For example, in section B9/R (Chapter 4), the sediment grades upward from a matrix-dominated diamicton to a clast-dominated diamicton to a massive unit of well-sorted silt at the top. [2]

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[2]These sections were logged and interpreted before the study by Eyles and Eyles (1983) was published. The similarity between some of the sections near the eastern side of the Foothills and those described by Eyles and Eyles (1983) at the Scarborough Bluffs are sufficiently similar to warrant further investigation of the Foothills exposures.

LACUSTRINE-MANTLED GROUND MORAINE

Within and along the eastern slopes of the Foothills are several areas where lacustrine sediments mantle or mask the underlying deposits but fail to produce a distinctive landscape. In all of those areas the substrate was till so the designator of lacustrine-mantled ground moraine was adopted, although there is no *a priori* reason why lacustrine mantles were not superimposed on other deposits. Lacustrine-mantled ground moraine is particularly common along the eastern slopes of the Foothills where deglaciation resulted in the formation of ephemeral ice-marginal lakes. Where the resulting lake deposits are sufficiently thick to mask the underlying deposits, the area was mapped as a lacustrine plain, but where the lacustrine deposits are thin, intermittent and fail to mask the topography of the underlying material, they were mapped as a lacustrine-mantle. Exposures in these deposits are rare. The deposits usually occur at low elevations and in areas of low relief, thus, they are usually very wet. In fact, the presence of extensive tracts of organic terrain is the prime indicator of a substratum of lacustrine sediments. The occurrence of such deposits was verified by hand-augering where possible.

In places along the Brazeau River, more than 5 m of lacustrine sand, silt and clay overlying till were measured. Large clasts incorporated in these deposits probably represent dropstones or re-worked debris-flow material that cascaded into the ponds. In many sections the basal material contained such a high coarse clast content that it was difficult to differentiate it from the underlying till. The significance of this unit is that it identifies areas where local ponding occurred during deglaciation. The presence of these deposits in some tributary valleys in the Foothills south of the Brazeau River valley supports the contention that ice downwasted in the tributary valleys while ice within the main valley was still active. The extensive lacustrine mantle along the eastern slopes of the Foothills indicates that ice-marginal lakes formed between the Foothills and an ice mass to the east and northeast (the Athabasca Lobe). The variability of the texture of these lake deposits, and the lack of varves, suggest rapidly fluctuating depositional environments. The frequent inclusions of dropstones indicate that much of the ice floating in these lakes was carrying debris or that debris-flows from adjacent ice surfaces were entering these water bodies.

It was commonly found that the glaciolacustrine deposits graded upward from ice-contact material containing a high percentage of clasts to well-sorted fluvial sand and lacustrine silt near the top. Presumably, this reflects the progressive removal of the ice from the vicinity of the ponds.

3.5 RECENT (HOLOCENE) LANDFORMS

As stated in the introduction of this study, the main focus is on Pleistocene landforms. Consequently, the Holocene alteration of the landscape has been de-emphasized. Along the stream valleys and in the mountains it is difficult to ignore some Holocene features as they overlie, mask, or dissect the Pleistocene deposits. In the high alpine areas the Holocene moraines and rock glaciers have a more direct bearing on the study as they relate specifically to the glacial history of the area. For these reasons, a selected group of landform units has been mapped and interpreted.

3.5.1 FLUVIAL LANDFORMS

STREAM CHANNELS, BARS, FANS AND FLOODPLAINS

A single category (streams) was used to identify all the landforms created by recent fluvial activity. Erosional and depositional landforms have not been differentiated but the topographic and morphologic evidence presented on the Landforms Map should enable the reader to subdivide this unit. Large alluvial fans mask sections of the Pleistocene deposits within and along the front of the mountains. Most of these are stabilized and present streams are incised into their surfaces. At least two, and frequently three, tephra layers are usually found within the upper metre of these deposits. Westgate and Dreimanis (1967), Luckman and Osborn (1979), and Jackson (pers comm, 1983) identified similar ash units as Mazama (6,600 B.P.), St. Helens Y (3400 B.P.) and Bridge River (2600 B.P.) tephra. This suggests that much of the fan aggradation occurred during early Holocene time. At the exit of the Southesk and Brazeau Rivers from the mountain front the streams occupy a channel with a relatively wide floodplain but downstream the valleys become deeply incised into the Pleistocene deposits and excellent stratigraphic sections were found on both sides of the rivers. Up to 40 m of unconsolidated material have been measured along these sections of the rivers. The Southesk River differs from

the Brazeau River in that it is incised into bedrock from the mountain front to about 3 km upstream of its confluence with the Brazeau River where deep sections of Pleistocene deposits are exposed. This indicates that the main Brazeau River valley was more deeply eroded than were the smaller ones such as the Southesk River valley. If this was largely due to glacial erosion, as the valley shape and the geologic structure of the area suggests, the Brazeau Valley was the dominant ice corridor along this section of the mountain front.

With the exception of the high terraces along the valley sides the Brazeau River remains within a deep canyon until beyond the eastern ridges of the Foothills. Through the Foothills the steep valley walls are very active and moderate summer storms have been observed to trigger debris-flows and rockfalls. To the east of the Foothills the valley morphology changes abruptly downstream from the esker complex discussed earlier. The valley widens and tree-covered slopes replace the steep unstable slopes found further up-valley. The channel pattern becomes progressively more braided downstream from this point to the Brazeau Reservoir. A probable major reason for the change in channel pattern is the removal of topographic constraints on the valley downstream of the Foothills. Another contributing factor may be the influx to the bedload of sands and gravels from the thick sequences of glacial, glaciofluvial and glaciolacustrine material through which the river has incised east of the Foothills.

3.5.2 MASS MOVEMENT LANDFORMS

SCREE APRONS AND SCREE FANS

These are extensive within the mountain portion of the study area and are particularly associated with the limestones of the Front Ranges. At higher elevations (above 2250 m) some of the thicker accumulations of scree, in east or northeast-facing cirques, have mobilized to produce rock glaciers. Spectacular rock glaciers occur in tributaries of the Brazeau River valley upstream of the study area. It is noteworthy that the rock glaciers tend to be concentrated in valleys with calcareous bedrock and are sparse or nonexistent in valleys dominated by older metamorphic rocks to the west.

Scree appears to be readily available in both areas. The explanation for this phenomenon is not immediately obvious, but the weathering properties of the rock types, and the characteristic sizes and shapes of the boulders within the scree deposits, probably warrant investigation.

Holocene landslides are evident in the mountains of the study area and further to the west. Oversteepening of the valleys by glacial erosion during the Quaternary has created an environment vulnerable to landslides, and seismic activity has probably been responsible for triggering many of the slides. Mr. R. Wetmiller, of the Earth Physics Branch of Energy Mines and Resources, supplied data indicating that each year there are several low-intensity earthquakes in the study area. These may be capable of triggering landslides and avalanches. A summary of the events recorded between Latitudes 52 and 53 degrees N and Longitudes 116 and 118 degrees W during the years 1967-77 indicates that local earthquakes range up to 4.6 on the Richter Scale, with a modal intensity of 2.6. Eleven of the 32 events registered higher than 3.0. Wetmiller also stated that the Laramide Faults in this area have been relatively active over at least the last decade and the level of activity appears to be increasing south of Latitude 54 degrees North. Triggering of large-scale mass-movement may be expected from these seismic events.

3.5.3 GLACIAL LANDFORMS

MORAINES

Many of the mountain valleys have well-defined end moraines and lateral moraines, usually a kilometre or less downsteam from their cirque headwalls. These moraines appear to be similar to those described by Luckman and Osborn (1979), two nested end moraines are most common, although three are sometimes found. All Holocene moraines are high in the valleys, usually at elevations above 2100 metres. No Neoglacial moraines were found in the south fork of the Brazeau River valley to the east of Nigel Pass. The north fork heads at the present-day glacier on Mount Brazeau. The Neoglacial end moraines in the north fork valley are irregular and difficult to interpret

because of varied local topography and the interaction of several distinct lobes of ice within this part of the basin. Nevertheless, the most distal end moraine is within 3 km of the present ice front, near the threshold of a hanging valley. The Holocene moraines were not investigated on the ground so that details regarding the degree of weathering, slope angles and the soil development on these features are not known.

AEOLIAN LANDFORMS

No thick dunes or loess deposits comparable to the Brule Lake deposits in the Athabasca Valley were observed within the study area. However, in most of the soil pits excavated in the eastern Foothills and the adjacent plains the **Ah** horizon had a high silt content indicating the presence of loess. Two soil pits (Sections B17/R and B15/R) excavated within the Western Alberta Plains revealed 10 and 36 cm respectively of loess overlying a buried organic horizon. Terrestrial plant fragments, the absence of fine silt and clay, the absence of laminae and the discontinuous nature of the deposits provided the basis for classifying these deposits as aeolian rather than lacustrine. Loess sediments, probably of Holocene age, mantle most of the area but the deposits have largely been incorporated into the solum and rarely form identifiable landforms.

4. STRATIGRAPHY AND SEDIMENTOLOGY

4.1 INTRODUCTION

Stream cuts, a few roadside exposures, borrow pits, the exposures created during the creation of oil well sites, seismic lines and a few well-log reports are the main sources of stratigraphic evidence discussed in this study. The cutbank sections along the watercourses are not so well exposed as those in southwestern Alberta, largely because of more rapid erosion and mass-movement rates stimulated by the greater annual moisture surplus of the north-central foothills. The exposures that do exist are concentrated in two main areas; just to the east of the mountain front and immediately east of the Foothills. Selected sections were measured and described in both of these areas and the information provides part of the base for the deglaciation model suggested later in this study. The principal objective of the stratigraphic investigations was to complement the identification and interpretation of the landform assemblages in the study area. In addition, a search was conducted for buried or relict (i.e. pre-Late Wisconsin) glacial deposits which might provide evidence for earlier glacial events than those responsible for the main suites of landforms. The definitions of some of the stratigraphic terms, such as 'till', are the subject of considerable debate. It is the opinion of this author that much of this debate persists because of failure to recognize that these terms must be defined within an explicit (or assumed) context. The rigor of the definition is therefore a function of the context. In this study the objective is to identify the dominant geomorphic agent. A sedimentologist observing the same deposits would choose to use a more precise set of terms because the sediments would then represent the principal line of evidence. Because sedimentology and stratigraphy play only a supportive role in this study, the deposits are described with less precision (but not less accurately!). For example, if a unit of glaciofluvial sediment could be demonstrated to have been thrust slightly by subsequent glacier action, or, if it collapsed slightly during the wastage of adjacent ice it would be described as a glaciofluvial deposit for the purpose of this study.

In this study a diamicton is interpreted as being a till if it contains material which is not of local origin and at least some of the clasts display evidence of having been

worked by ice. It is recognized that some of the deposits interpreted as till may have been transported from an ice surface by mass movement processes but the consequences of an error of this sort would be insignificant in terms of the objectives of this study. Figure 4.1 shows the locations of the major sediment sampling and description sites discussed in this chapter.

4.2 STRATIGRAPHIC SECTIONS AND SEDIMENTOLOGIC INTERPRETATIONS

4.2.1 SECTIONS BETWEEN THE FRONT RANGES AND THE FOOTHILLS:

SECTION B7/L

The Brazeau River in this location has truncated a drumlin and the upper surface of the section is approximately accordant with similar landforms to the southeast of the river. These are the highest relief features within this segment of the valley. Ten and one-half metres of diamicton were measured overlying poorly sorted and stratified fluvial or glaciofluvial deposits. The till is light buff to grey-buff in color, poorly indurated, and contains rounded to sub-angular clasts ranging up to 1.8 m in diameter. The modal clast size is approximately 10 cm in the coarser fraction and the fine fraction is dominated by sand and silt with very little clay. Limestone and sandstone are the dominant lithologies with quartzites also common. Fractured quartzite clasts and facetted limestone clasts suggest that the diamicton unit is a till. This interpretation is augmented by the fact that the upper surface of the unit is drumlinized. The lower contact of the unit terminates with a sharp transition to a fluvial/glaciofluvial unit. The lower unit coarsens upward and the upper few metres are comprised of very coarse material with boulder diameters frequently in the 20 – 30 cm diameter range.

In general the lower unit consists of poorly to moderately sorted coarse gravels in a silt-sand matrix. Evidence of stratification was not visible but cut-and-fill structures were observed. The depth of this unit is approximately 10 to 15 m but an apron of talus which mantles the lower portions precluded an accurate measurement. Extrapolating from smaller exposures, near river-level downstream, it is assumed that the

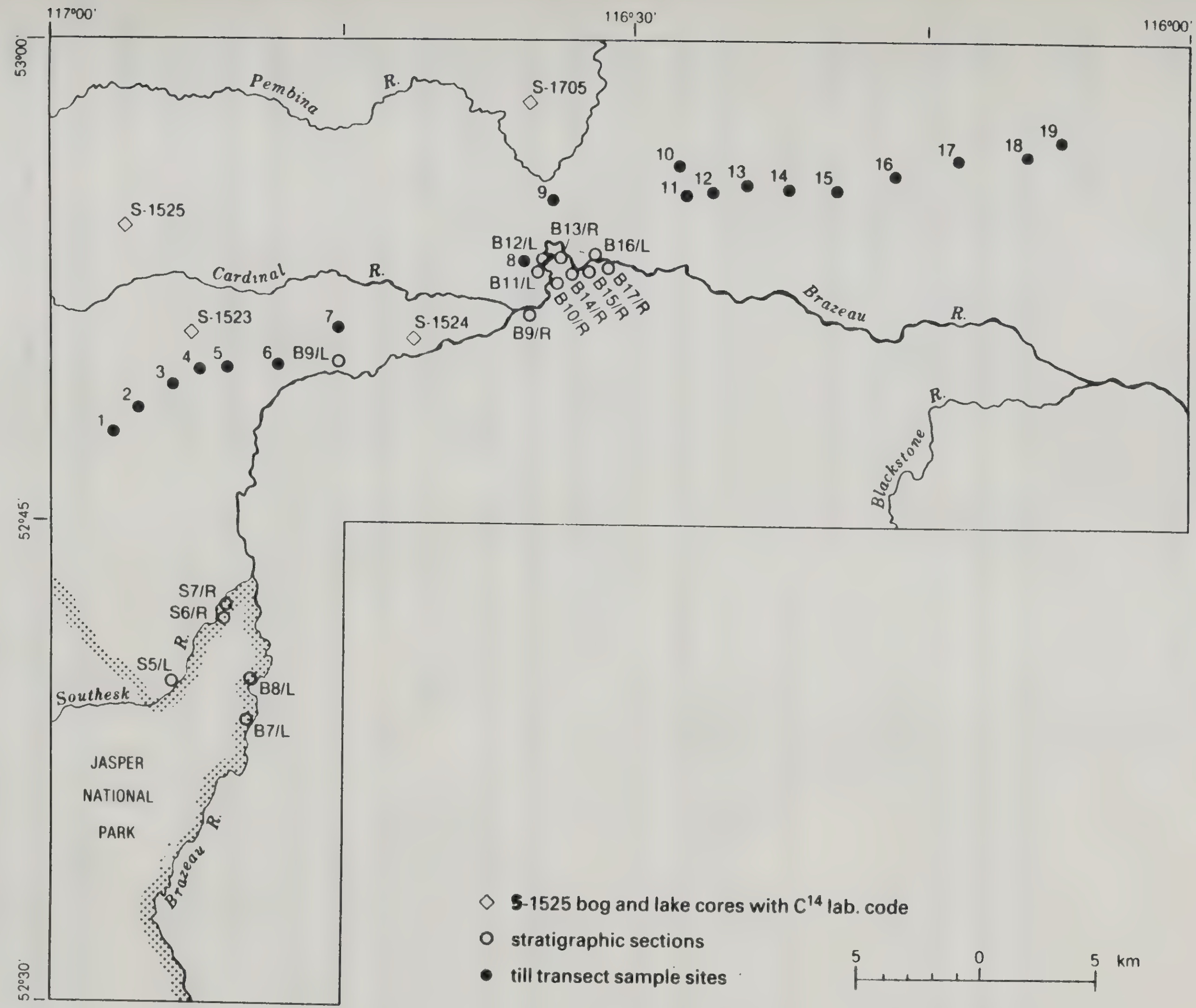

Fig. 4.1 Sample Locations

fluvial/glaciofluvial unit directly overlies bedrock at an elevation a few metres above present river level. The definite coarsening upward which was observed in the exposed portion of the lower unit, and the cut-and-fill structures, suggest that this unit is part of an outwash train deposited in front of an advancing glacier.

SECTION B8/L

The best exposures found within the study area lie on either side of the Brazeau River a few kilometres upstream from the mouth of the Southesk River. The stratigraphy of these sections is relatively simple and consistent so the description offered below should be considered as representative of the local area. Figure 4.2 shows the general sequence of units which occur at this section.

Unit **a** consists of 1 to 1.5 m of grey-buff diamicton. It is less indurated than the underlying units and forms slopes of up to 70 degrees as illustrated in Fig. 4.2. Rounded to angular clasts of limestone, sandstone and quartzite up to 80 cm in diameter are found but the modal size of the coarse fraction is approximately 30 cm. Broken and facetted clasts, as well as the planar upper surface of this unit, indicate that this is a till unit. Sand lenses are common and many of the coarse clasts appear to have been previously rounded by fluvial action. The contact at the base of unit **a** is irregular and thin (less than 50 cm) lenses of distorted beds of sand and gravel occur at the base of the unit. It is important to note that although these lenses occur at a preferred level, they are neither continuous nor do they always occur at the same level. This indicates that these sand inclusions were more probably formed by englacial or supraglacial meltwater than by a separate non-glacial event. Haldorsen and Shaw (1982) describe the genesis of such interstratified deposits and explain how they are an important diagnostic indicator of melt-out till.

Unit **b** consists of approximately 10 m of buff-colored diamicton with a matric which is finer textured than unit **a**. Most of the clasts are angular to sub-angular but rounded clasts were also noted. The size of the coarse fraction ranges up to 50 cm with a modal size for the coarse fraction being about 25 cm. Slope angles up to 75 degrees were measured, indicating a greater degree of induration. A thin (50 cm) band of water-rounded cobbles in a poorly-sorted matrix is found at the base of unit **b**. This

band is discontinuous. In some sections it could not be found while at other locations it was replaced by coarse gravel or sand. It is generally present in the section described for the left side of the river but was not observed on similar sections directly across the Brazeau River.

Unit **c** is about 12 m thick and, except where separated by sand and/or gravel lenses, it is distinguishable from unit **b** only by the fact that it is somewhat more indurated. Color and texture remain the same as for unit **b** but slope angles are steeper within unit **c**. There appears to be a concentration of larger clasts in the lower 1 to 2 m of the unit.

Unit **d** is a highly indurated mass of ice-contact drift or very poorly sorted outwash. This unit is typically 15 to 17 m thick. Slope angles up to 80 degrees were measured and when the exposure is dry vigorous blows with a hammer are required for the removal of material. Modal clast size is 15 cm for the coarse fraction but commonly clasts up to 35 cm in diameter were observed. The matrix is highly variable but tends to be comprised of a high percentage of sand sized particles. Directly underlying unit **d** is shale bedrock which is less resistant than the material of unit **d**.

No weathering zones, oxidation zones (except for the top of unit **a**) or obviously nonglacial facies were found within the section. Unit **d** seems to be pro-glacial outwash or ice-contact stratified drift which was deposited in front of the advancing ice. Unit **c** may be a till that was deposited from the base or sole of the glacier and unit **b** may be englacially transported debris which was deposited either by melt-out or the lowering of supraglacial material. If this interpretation is correct, one would expect to find striated or facetted clast and bullet-shaped boulders in unit **c** but no such evidence was observed. Alternatively, units **b** and **c** could be genetically similar and the gravel unit may be a relatively insignificant local phenomenon. A third alternative is that the gravel unit represents a lag deposit indicating a hiatus in the depositional record and unit **c** represents an earlier advance than unit **b**. Unit **a** is tentatively interpreted as an ablation till or a flow diamicton in that it contains a relatively high percentage of coarse clasts, most of which show evidence of having been fluvially worked. In addition, unit **a** differs in color from the underlying material. Again, it might be argued that this unit represents a subsequent glacial event but there is no clear contact between this unit and the underlying

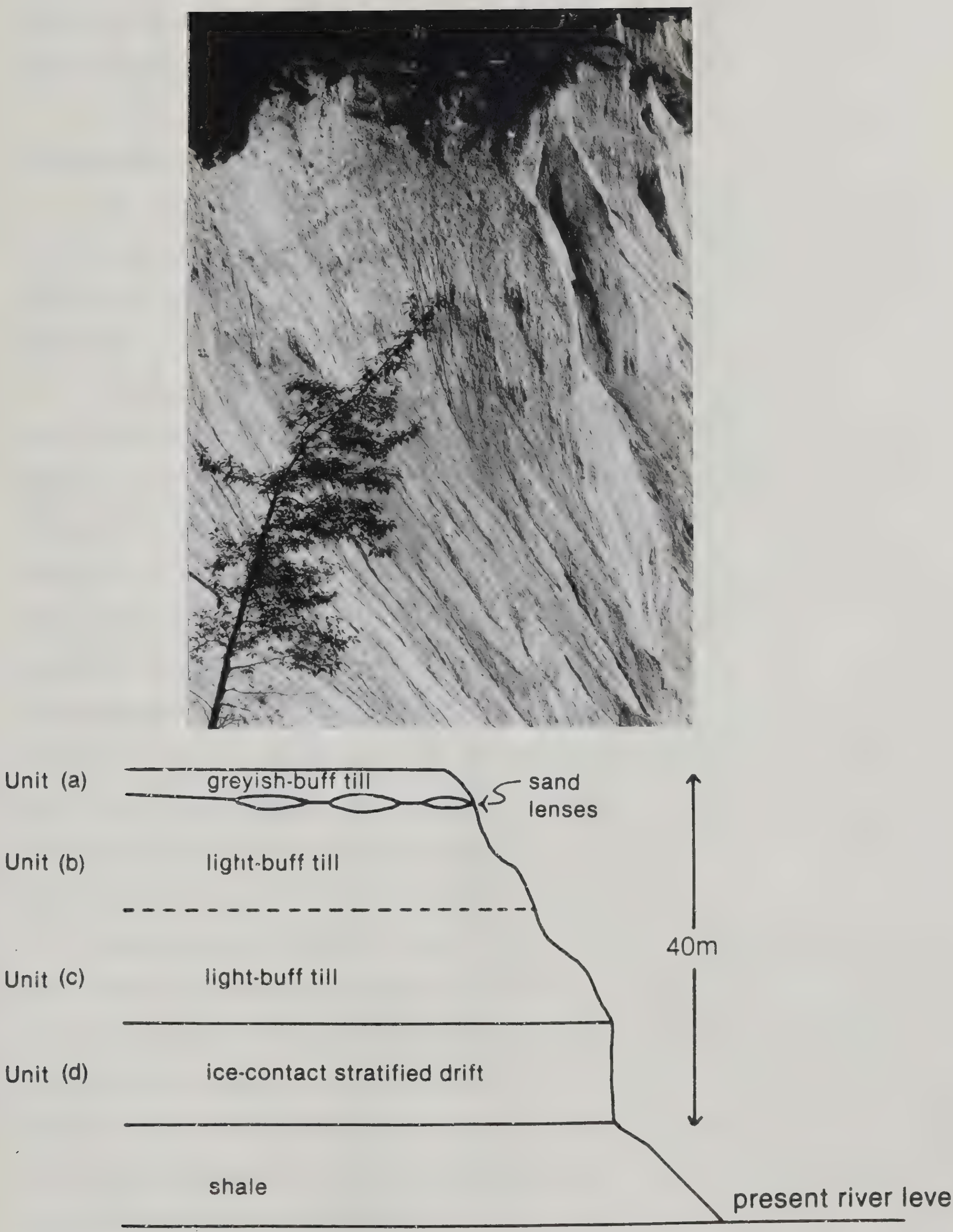

Fig. 4.2 Stratigraphy, Section B8/L

units nor is there evidence of a weathering horizon or an erosional surface which would tend to support this hypothesis.

SECTION S5/L

This section is located immediately east of the exit of the Southesk River from the Front Ranges. The stratigraphic sequence is closely comparable to that found at section B7/L and the two sections are located in similar positions within the Brazeau River valley.

An upper unit comprises 8 m of buff-colored till and this overlies 17 m of poorly sorted coarse gravel. The till unit contains limestone, sandstone and some quartzite clasts, angular to sub-rounded in shape. The larger clasts are typically in the 18 - 20 cm size range although some range up to 50 cm in diameter. The fine fraction of the till unit is a mixture of sand, silt and clay and is moderately indurated. The underlying gravel unit is moderately to poorly sorted and poorly stratified. Clasts are rounded to sub-angular (mostly sub-angular) with facetted surfaces. The modal size of the clasts is 15 cm, ranging up to 35 cm in diameter. This unit is sufficiently indurated that a hammer is required to remove the samples when the material is dry. Visible calcium carbonate on many of the clasts suggests that carbonate cementation is the cause of the induration. Lithologies are dominated by limestone, sandstone and quartzite. Sandstone bedrock was observed at the base of the section directly underlying the lower unit.

The stratigraphic record at this location appears to indicate a glacial advance which deposited a till unit on top of a pro-glacial outwash unit. The lack of obvious coarsening-upward in the outwash unit may be due to truncation of the upper portion of the gravel unit during the glacial advance or, alternatively, the unit may indeed coarsen-upward but the evidence is masked by the fact that the unit is very poorly sorted and a coarsening-upward trend could only be statistically demonstrated with a large number of clast measurements.

SECTION S6/R

The upper unit **a** at this section (Fig. 4.3) consists of a 1 m thick band of fine sand, silt and clay with no gravel or larger clasts. Beneath this lies unit **b**, 2 - 3 m of coarse,

poorly sorted gravel comprised of sub-angular to sub-rounded clasts which range up to 1 m in diameter. The modal size of the clasts in the coarse fraction is approximately 20 cm. Cementation of this unit is moderate but the face can be broken with a shovel, unlike the more indurated, but otherwise comparable unit described for section S5/L. Cross-stratification is evident within this unit. A wavy contact separates unit **b** from unit **c** with troughs approximately 1 m deep cut into the till. Unit **c** consists of 10 m of buff-colored compact diamicton containing clasts up to 1 m in diameter. Slope angles up to 75 degrees were measured on this unit. Beneath the diamicton lies 15 m of coarse gravel with irregular sand inclusions. Modal clast size in the coarser material is approximately 15 cm with the largest clasts up to 30 cm in diameter. The matrix material is clayey sand. Frequent inclusions of unweathered shale suggest that a large amount of local bedrock has been incorporated in this unit. Unit **d** is subdivided into upper, 7 m thick, and a lower, 8 m thick, sub-units on the basis of color. There appears to be an erosional surface between them with gravels of the upper bed filling troughs in the lower sub-unit. The texture of the two units appears to change very little but the upper unit contains darker bands whereas the lower sub-unit is more uniform in color. This may suggest that more local bedrock is incorporated in the upper material, or perhaps that a water table once lay above the lower sub-unit.

SECTION S7/R

At this site (Fig. 4.3) thin deposits of silt (assumed to be loess on the basis of its texture compaction and lack of bedding structures) up to 40 cm deep were found overlying the upper diamicton unit. These are not indicated on Fig. 4.3 because of the irregular depth and discontinuous nature of the material. The upper unit **a** of this section consists of a 1 - 1.5 m thick mantle of light grey diamicton. The upper diamicton contains few angular clasts and is poorly consolidated (breaks easily with a shovel). Silts and sands predominate in the matrix and sandy inclusions were observed. The contact between the upper and lower diamictons is distinct but irregular. In places poorly sorted gravel was found along the contact. The lower unit (unit **c**) is a coarse textured bed incorporating a high percentage of sub-angular and sub-rounded clasts with a modal size of 15 cm, ranging up to 80 cm in diameter. Striations were found on some of the

limestone clasts. The unit is highly indurated and, when dry, a hammer is required to remove samples. In a bed considered to be the stratigraphic equivalent of this unit, but on the opposite side of the river, large (>3 m) irregularly-shaped inclusions of local bedrock were found incorporated into an otherwise much finer-textured sediment. Eroded into part of unit **c** is a unit of gravel which pinches out downvalley. The upvalley end of this unit is approximately 2.5 m thick and is comprised of coarse gravel (modal size of 20 cm). The surface morphology of this unit is mapped as a terrace on the Landforms map. The internal composition of the unit was described earlier as unit **b** of S6/R which can be traced continuously along the length of the exposure. It does appear, however, that the cross-stratification is more pronounced toward the downvalley end of the exposure and the texture of the gravel tends to fine in the same direction. The underlying diamicton unit is in turn underlain by coarse, poorly-sorted gravel (unit **d**) which is thinner at this location than it is at Section S6/R. There is no evidence of the change in color which was observed at the southern Section S6/R. The unit is partially masked with debris between the two exposures so it is not known where the color difference begins. Possibly the upper portion of unit **d** of section S6/R pinches out as the unit at Section S7/R end is similar to the lower portion of unit **d** (see Figure 4.3). At S6/R unit **d** reveals a definite coarsening upward of the sediments. In fact, the transition to the overlying diamicton unit, interpreted as being a till, is marked by an abrupt transition from 30 cm diameter, sub-angular boulders to diamicton. The unit is well stratified and moderately well sorted but near the downvalley end of the exposure an inclusion of well-bedded fine gravel and sand is inclined upward at an angle of 8 degrees. This appears to have been due to differential compaction or, possibly, ice-thrusting during glaciation. A more striking example was found on the north side of the river slightly to the east of this location but in the same stratigraphic unit (See Fig. 4.4)

The stratigraphies of S6/R and S7/R are interpreted as follows: On the basis of landform evidence, sedimentary characteristics such as a relative lack of sorting, fractured and facetted clasts and stratigraphic association, the diamictons are interpreted as till. Till unit **a** is probably an ablation or flow till which was draped over the surface of the major till unit (unit **c**). The indurated nature of till **c** suggests that it was deposited subglacially, compacted and later cemented by calcite. Friable bedrock inclusions support

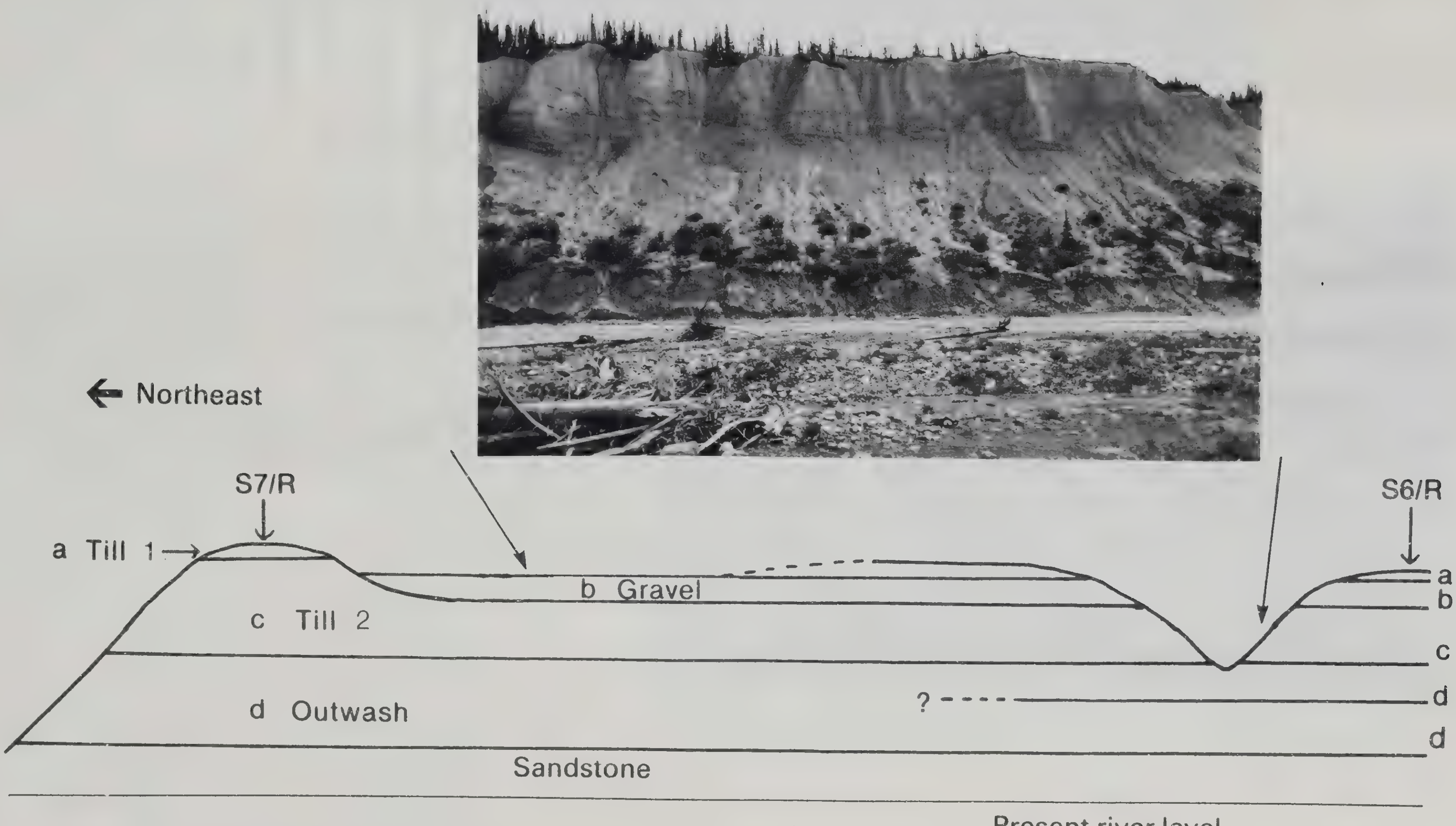

Fig 4.3 Stratigraphy, Sections S6/R and S7/R

Fig. 4.4 Collapsed or ice-thrust sediment exposed within Southesk River Valley

the argument that these may represent englacially-transported sediments as it is doubtful that consolidated material could withstand the shear stress which would be expected with subglacial sediments. It is also unlikely that such materials would have been preserved if they were exposed at the surface of a glacier and subsequently lowered during ablation.

The coarsening-upward of the unit **d** sediments suggests that these are probably outwash or ice-contact deposits formed during a glacial advance. The dark color of the upper portion of S6/R unit **d** indicates that some of the sediment was probably derived from local sources. Black calcareous shales (Upper Cretaceous) outcrop immediately east of the mountain front, only a few kilometres upvalley from these exposures (Holter and McLaws, 1974). The significance of the color change within the gravel unit is not known.

All of these sections lie within the broad depression between the Front Ranges and the Foothills. This topographic setting would have probably produced divergent flow in the Brazeau Valley glacier as it emerged from the more confined segment of the valley and expanded into a piedmont lobe. The orientations of ice-moulded landforms in the area indicate that divergence did occur at least during the later stages of active ice flow during the last glaciation. Divergence of the ice would reduce the ice thickness, disperse the sediment load over a wider area, and would therefore reduce the erosional potential of the glacier, thereby increase the probability of preserving the pre-existent deposits. Therefore, evidence for multiple glaciations would be most likely to exist in such environments. No unequivocal sedimentary evidence of multiple glaciations was observed in these sections insofar as the till units may be explained on the basis of facies related to a single glacial event. A generalized description of the sediments associated with this advance is that the glacier advanced from the mountains depositing a unit of outwash gravels or ice-contact stratified drift. A unit of approximately 10 - 20 m of glacigenic sediments, primarily till, directly overlies the lower unit. In some sections the till unit contains evidence of melt-out or ablation (or flow) facies as well as a basal facies. The till unit may extend to the surface, or it may be overlain by a thin (<2 m) unit of aeolian, lacustrine or fluvial material.

4.2.2 SECTIONS WITHIN THE FOOTHILLS:

Very few exposures were found where a complete stratigraphic sequence above bedrock could be observed. Along the north wall of the Brazeau River valley one exposure was found (B9/L, Fig. 4.1) where approximately 2 m of diamicton directly overlies bedrock.

SECTION B9/L

This is the only section found within the Foothills zone of the study area in which glacial sediments were exposed and the lower contact of the diamicton unit could be determined. This thin diamicton overlies shale bedrock. Generally the thickness of this unit ranges from 1.5 - 2 m and it is comprised of rounded to sub-rounded clasts in a silty sand matrix. Clasts range up to 30 cm in diameter but modal diameters are 4 - 5 cm. The

diamicton is dark grey in color below the soil B horizon and is poorly consolidated. Limestone and sandstone are the dominant coarse clast lithologies but quartzite and siltstone are not rare. Some broken clasts with sharp edges were observed but most of the material appears to have been rounded. The matrix is silt or sand dominated which suggests that the finer fraction has been removed. This material is interpreted as a till which has undergone some alteration by water. it is probably of supraglacial origin and is likely an ablation till or a flow deposit.

The Thunder Lake esker compex has been dissected by the Brazeau River and related riverbank exposures clearly show that the esker sediments directly overlie bedrock. It is the lack of till sediments through this glaciofluvial section of the study area which precludes the direct stratigraphic comparison of the tills along the west side of the Foothills with those to the east of the Foothills. If, as some researchers (Roed 1968, 1975, Reimchen and Bayrock 1977, Boydell 1978, Rutter 1978, Stalker 1978) believe, the Foothills were only glacerized by local valley glaciers during the last major glaciation, the terminus of that advance would probably have been located near the eastern limit of this complex.

SECTION B9/R

Near the eastern edge of the Foothills section B9/R (Fig. 4.1) revealed 8.3 m of glacial deposits. This section proved difficult to interpret because the lower 4.9 m of material displayed characteristics quite different from any material observed in the sections adjacent to the mountain front. Although this unit is tentatively designated as a till, the sandy nature of the matrix, the inclusions of sorted and stratified material and the loose nature of the sediments suggest that the material was deposited in a very wet environment, possibly into standing water. The site is located along the south side of the Brazeau River valley at its intersection with a tributary valley which presently contains Moosehound Creek.

The upper unit consists of 3.4 m of poorly consolidated, well-sorted and laminated silts with scattered water-rounded clasts up to 2 cm in diameter. The larger clasts are commonly found within rhythmites of fine sand, silt or clay and are therefore interpreted as dropstones. Below the upper unit lies 0.7 m of oxidized (reddish-brown),

fine-textured diamicton. Modal clast size is approximately 1.5 cm in the upper portion of the unit and increases to 2 - 3 cm in the lower portion but clasts up to 12 cm in diameter can be found throughout the unit. Near the top of the unit 10 per cent or less of the exposed face consists of coarse clasts whereas near the bottom, 20 per cent or more of the face is composed of coarse clast surfaces. The diamicton is poorly consolidated (slope angles are typically up to 32 degrees) and the larger clasts tend to be sub-angular to sub-rounded with rounded clasts not uncommon. The texture of the matrix is sandy-silt.

A second diamicton unit 4.2 m thick underlies the upper unit. The contact is marked by a distinct color change to dark grey or blue-grey, increased induration (slopes as steep as 40 degrees) and a slight increase in the frequency of sub-rounded to rounded clast shapes. Iron staining was observed on many clasts. The matrix is remarkably sandy considering the compact or indurated nature of the unit. Approximately 1.3 m below the top of this unit the incidence of coarse clasts increases from 10 to 15 per cent of the face to about 50 per cent. The modal clast size is about 3 cm in diameter but clasts with diameters up to 20 cm in the upper portion, and 70 cm in the lower portion of the unit were observed. Near the base of the lower unit the matrix becomes finer and the color of the unit becomes darker, possibly reflecting a higher incidence of locally derived shale. The incidence of coarse clasts gradually decreases beneath the coarser band and in the lower section less than 10 per cent of the face is comprised of exposed coarse clasts. The lower diamicton rests directly on a unit of black calcareous shale. The interpretation of this section is as follows: The lower diamicton is interpreted as till on the basis of its composition and the inclusions of fractured clasts. The gradual transition to the middle unit (diamicton) indicates that no fundamental changes occurred in the environment during the time of deposition. The weakly stratified and poorly sorted nature of the middle unit, together with the concentration of clasts in the lower portions of this unit, suggests a saturated ice marginal environment. Obviously this is a waterlain till, as described by Dreimanis (1976), or a subaqueous flow deposit from an ice surface into an ice-marginal pond. In any case, it seems clear that a glacier margin must have been present at the time of deposition in order to accumulate this material along the valley wall and drainage down-valley must have been at least locally obstructed in order to allow

subaqueous flows to exist in an environment where local topographic characteristics would dictate otherwise.

The upper unit is thought to represent a glaciolacustrine unit on the basis of the texture and stratification of the sediments and the larger clastic inclusions are interpreted as ice-rafted material (drop-stones). This sequence represents a continuous transition from glacial, through glaciolacustrine, to lacustrine environments. sediments which seem to display evidence of both environments.

4.2.3 SECTIONS ALONG THE EASTERN SLOPES OF THE FOOTHILLS

In addition to the piedmont area east of the mountain front, one might have expected to find superimposed glacial deposits east of the Foothills where the topographic constraints provided by the Foothills ridges ceased to be significant. The stratigraphic sections measured along the Brazeau River valley in this area were much thinner than those adjacent to the mountain front. Tills which were lithologically and texturally similar to those found further west were identified in the adjacent upland areas, but the sedimentary sequences found in the sections along the Brazeau River valley were significantly different. In the eight sections measured along the Brazeau River east of the Foothills the stratigraphy was complex and difficult to interpret. Till, which comprises a major portion of the Pleistocene deposits within the study area, was completely absent in two sections (B11/L and B13/R) and comprised a major portion of only the three easternmost sections (B15/R, B16/L and B17/R). Moreover, some of the diamictons which have been tentatively described as till frequently display characteristics that are atypical of tills. For example, large clasts tend to increase in frequency in the lower portion of some units, sorted and stratified inclusions are common throughout the deposits and the fine silt and clay fractions appear to have been partially removed. A useful analog for these deposits is found in the type IV sediment flow deposits described by Lawson (1979, p. 56) from the western terminus of the Matanuska Glacier.

SECTION B10/R

Section B10/R is located east of the Cardinal Hills, which provide the last major topographic constriction to the Brazeau River valley, and just upstream of the Big Bend

of the Brazeau River. The upper unit of this section consists of 1.2 m of buff-colored lacustrine silt and clay. No dropstones or coarse sediment bands were observed in this unit. Directly beneath the upper unit lies nearly a metre (0.7 m) of well-sorted, stratified sand and gravel with frequent evidence of cross-bedding. The texture of these sediments is variable with some clasts up to 15 cm in diameter but with a modal size of approximately 1 cm. The upper portion of the sand and gravel unit is unconsolidated and frequently forms a niche in the exposure. Underlying this is nearly a metre of poorly-sorted, coarse sand and gravel. Clasts are sub-angular to rounded and range up to 20 cm with a modal size of 6 to 7 centimetres. Below, and grading into the sand and gravel unit, are 3 m of light-buff colored diamicton containing both water-rounded and angular clasts (modal size 3 to 5 cm, ranging up to 80 cm in diameter) in a clay-silt matrix. The unit is compact but not indurated and the texture of the coarse fraction increases with depth, particularly in the lowest metre of the unit. It appears that the sediments were deposited in a saturated condition, possibly a subaqueous environment, and the larger clasts settled during transport. Weathered sandstone directly underlies the diamicton unit. It is suggested that the lower unit at this section represents a waterlain till or ice-marginal diamicton (flow deposit) which was deposited under wet conditions. The overlying sands and gravels are most probably of glaciofluvial origin and the gradational boundary between the sands and gravels and the silts implies that the silts probably represent glaciolacustrine deposition.

SECTION B11/L

Section B11/L is approximately 1 km downstream from B10/R and on the opposite side of the river. The upper 5.9 m of this section consist of well laminated silts and clays with laminated bedding structures clearly visible. The rhythmites include sand, silt and clay beds. No dropstones were observed. The basal part of the upper section consists of an abrupt transition to horizontally bedded gravel and coarse sand similar to that found at section B10/R. Rounded to sub-rounded clasts ranging up to 10 cm in diameter, with a modal size of 1 to 3 cm, characterize this unit. Iron oxide stains are common on clast surfaces in the upper 0.9 m whereas the lower 1.1 m shows no evidence of staining and a much darker color. It is thought that this difference represents

a relict watertable level. Beneath the two metres of bedded gravel lies a 1.7 m thick transitional unit comprised of bands of well-sorted sand separated by thick (10 to 30 cm) bands of poorly sorted and poorly or non-stratified diamicton ranging in texture from gravel to clay. This unit grades downward into a massive unit of diamicton containing clasts up to 8 cm in diameter (modal size equals 1 – 1.5 cm) in a clay-silt matrix. The clast shapes are sub-angular to rounded with frequent fractured surfaces. This unit is moderately consolidated, forming slopes up to 41 degrees. No evidence of stratification or sorting was observed, nor was there any evidence of staining on clast surfaces. The exposed portion of this diamicton unit is 2.6 m thick but the lower portion of the section was mantled by mudflow deposits so that the total thickness could not be determined.

This section is interpreted as a continuous sequence of deposits from till, at the base, through a series flow and turbidite deposits to the upper unit of lacustrine sediments.

SECTION B12/L

This section is interpreted as a unit of till grading upward into a sequence of subaqueous flow deposits and turbidites and then into a lacustrine deposit. All of the boundaries between the major units were gradational and no evidence was observed of any hiatus in the sedimentary sequence.

Section B12/L is located on the west limb of the Big Bend of the Brazeau River. The upper 4.3 m of this section consist of weakly laminated, well sorted fine silt and clay which grades downward into 2.3 m of stratified coarse sand. Beneath this is 1.1 m of massive silt mixed with rounded clasts. A distinct boundary separates the silt unit from an underlying coarse gravel unit 3.5 m thick. Coal fragments and fragments of calcareous shale are included in the gravel sediments indicating that at least some of the sediments are of relatively local origin. Clasts range up to 20 cm in diameter but the modal size is about 8 centimetres. The matrix consists of poorly sorted, sandy silt with no apparent stratification. The lowest unit consists of a dark grey compact diamicton containing clasts up to 20 cm in diameter within a clay-rich matrix. The clay content increases and the color value decreases toward the base. These characteristics probably reflect the

incorporation of a larger proportion of the underlying black shale. Some rounded clasts occur but angular and sub-angular shapes predominate. None of the units of this section is consolidated except for the diamicton unit which contains enough clay to form slopes up to 40 degrees. These are at least 5 degrees steeper than the slopes formed by the overlying units. This section is interpreted as a till unit which may have originated from either melt-out or lodgement processes. Underlying this is an ice-contact deposit. The upper silt, sand and clay unit reflects a glaciolacustrine environment with improved sorting and a decrease in grain size. Such a sequence may represent the gradual infilling of an ice-marginal lake.

Once again the sediments observed at this section may be explained in terms of till grading upward into lacustrine deposits with intervening material which has the sedimentary characteristics of flow sediments or waterlain till deposits.

SECTION B13/R

Section B13/R is located on the east limb of the Big Bend. The upper 1.1 m consists of well-rounded pea-gravel and laminated coarse sand. This grades downward into 4.5 m of coarse, stratified sand free of pebbles which, in turn, grades into 1.2 m of banded fine sand, silt and clay laminae. Underlying the stratified units are 2.4 m of massive silty clay in which there is only faint evidence of stratification. The base of this unit is marked by an abrupt transition to a fine-textured diamicton with rounded to angular clasts up to 12 cm in diameter. Most of this unit consists of a dark grey matrix comprised of clayey silt. The measured portion of this unit is approximately 2 m thick and in addition there are 2 to 3 m at the base which are masked by thick colluvial deposits. The sedimentary characteristics of the diamicton unit do not furnish sufficient evidence to indicate the precise depositional environment but the large inclusions of fine-textured material suggest that local shales probably comprise much of the original material. The unit is tentatively interpreted as a till. The overlying unit represents a complex of glaciolacustrine and/or glaciofluvial sequences. A similar sequence is found approximately 1.5 km downstream at section B14/R.

SECTION B14/R The upper 0.7 m of this section consists of stratified and sorted sand

laminae with identifiable ripple cross-laminations indicating a direction of flow from about 060 degrees. This would indicate flows from ice sources on the Western Alberta Plains toward the southwest. Because measured paleocurrent directions were highly variable these data should only be regarded as indicating the general hydraulic gradient at the time of deposition. The unit grades downward into a massive sub-unit consisting of 5.7 m of fine sand, silt and clay which is light brown in color and includes no larger clasts. The base of this unit is marked by an abrupt transition to laminated deposits. This appears to be a transitional facies as the upper portion of the unit consists of banded sands and silts whereas the lower portion grades downward into gravel near the base. The transitional sediments are 0.5 m thick and overlie at least 4.5 m of coarse gravel (modal diameter of 4 cm) comprised of clasts as large as 30 centimetres. Sub-angular to rounded clasts can be found within a silty-sand matrix. This unit is moderately well sorted and stratified. Mudflow deposits mask the lower few metres of the section but no evidence could be found in rills or gullies to indicate the presence of any unit between the gravel and bedrock. The upper portion of this section indicates a relatively low-energy glaciofluvial and glaciolacustrine environment. The lower unit is either ice-contact materialor a subaqueous flow deposit.

SECTION B15/R

This section lies about 1 km downstream from B14/R but the stratigraphy is quite different. The upper 1.4 m consists of a loose diamicton comprised of clasts ranging up to 15 cm in diameter (modal diameter 3 cm) in a matrix of coarse sand and silt. Most clasts are rounded but angular and broken clasts are not difficult to find. Maximum slope angles approach 35 degrees. A sharp boundary separates the upper unit from 1.5 m of well sorted and stratified sand. Ripple marks were preserved and field measurements indicated a flow direction from the northeast. The base of this unit grades rapidly into 0.5 m of coarse gravel in a coarse sand matrix which directly overlies 13.6 m of diamicton. The upper portion of this unit is relatively fine textured, comprised mainly of a granular silt-clay matrix with clasts ranging up to 20 cm in diameter (modal size of 2 cm). Much of the coarser material has been water worked but broken clasts were found. Lower in the unit, and not separated by any observable boundary, the texture of the diamicton

becomes much coarser. Modal clast size increases to about 3 cm, the frequency of coarse clasts increases and the matrix becomes sandier. Near the base of the unit cobble sized (5 to 6 cm diameter) clasts become dominant and the matrix is primarily coarse sand and gravel. Many clasts retain percussion marks and striations. Coal and sharp-edged shale fragments are present. This unit directly overlies sandstone bedrock. In this section the lower unit is a till probably a waterlain till or a flow till. This is tentatively suggested on the basis of the coal and shale fragments. The clast concentration near the base of the unit indicates a very wet depositional environment. The overlying sand and gravel is obviously glaciofluvial in origin. The fining-upward of the sediments possibly indicates that the stream gradient was lowering during deposition or that the sediment source was retreating from the area. The section exposes an entire transect of the infilled channel so it is unlikely that the upward fining only affects lateral shifts of the stream channel or floodplain deposits. The upper unit is difficult to explain unless it is a flow-till or debris flow that blanketed the sand and gravel deposits.

SECTION B16/L

The upper unit consists of 4.5 m of dark-brown, fine-textured diamicton containing clasts up to 12 cm in diameter (modal size of 1 to 2 cm) in a dense silty clay matrix. Many of the clasts have been water-rounded but broken clasts are present and coal fragments are common in the lower portions of the unit, which suggests that this is probably a till. The base of the diamicton is marked by a sharp transition to a 10 cm thick layer of laminated clays with scattered, well-rounded clasts up to 3 cm in diameter. Beneath the clay band is a thick (5.4 m) unit of coarse textured well-rounded clasts in a sandy silt matrix. Lenses of sand are frequently included in the otherwise coarser material. Modal clast diameters are 4 to 5 cm in diameter but they range up to 12 centimetres. The texture of this unit fines upward. Coal fragments are found in the lower portion of the unit. Boulders as large as 1 m are observable and lenses of pea-gravel are interbedded with the much coarser material. Stratification and sorting are poor except in the lenses of finer material. Black calcareous shale underlies the lower unit.

The interpretation of this section is as follows: The thick lower unit is probably poorly sorted flow sediments or ice-contact stratified drift. The thin clay-rich unit and

the inclusions of stratified material probably indicate that sedimentation took place into an ice-marginal pond as the sorting is too poor for it to have been deposited in a large body of water. The top unit appears to be ablation till or a flow deposit which has undergone little re-working as evidenced by the preservation of fragments of coal in the deposit. Another interpretation is that the upper unit is a glaciolacustrine deposit. The coarser fraction represents ice-rafted material and the finer fraction represents lacustrine sediment from a very turbid water body.

SECTION B17/R

Section B17/R is the most easterly section that was logged in this study. Approximately 17.5 m of sediments overlie dark brown calcareous shale at this section. The upper 9.2 m are comprised of fine-textured diamicton. Clasts up to 15 cm in diameter (modal size of 8 cm) are scattered thoughout a massive dark brown matrix of silt, fine sand and clay. The clast size and the percentage of coarse clasts increase with depth and the lower 3 m show evidence of banding and stratification, each band being somewhat less than a metre thick. No coal fragments were observed in the coarser fraction which consists largely of sandstone, quartzite and limestone. The unit is not compact or cemented, and is easily broken by hand, but the fine texture of the matrix permits the exposed sediments to form slopes up to 41 degrees. Beneath this is a 8.5 m thick unit of moderately sorted and stratified sand, silt and gravel. The modal clast size of the coarser fraction is 5 cm but the clasts range up to 12 centimetres in diameter. Most of the clasts are water-rounded and no coal or softer lithologies were observed. The texture of this unit coarsens downward from a modal size of 2 to 3 cm in the upper portions to 5 cm near the base. This section appears to consist of outwash or pre-glacial gravel overlain by a fine-textured till or flow deposit. The lack of compaction of the diamicton and the sorting in the lower part of the unit suggest that the deposit probably originated from a supraglacial or ice-marginal environment.

SUMMARY

The previous descriptions illustrate that, in contrast to the sections near the mountain front, a glaciofluvial unit does not form the lowest bed of any of the sections

east of the Foothills except at section B17/L. In fact, no units were identified that could be directly traced or correlated between the areas east and west of the Foothills. The glaciolacustrine beds at the top of sections B10 through B14 are the most consistent feature but the sedimentary characteristics and thickness of these vary considerably. It is apparent that most of the sediments were deposited in a saturated, if not subaqueous, environment. This augments the landform evidence which indicates that ice-marginal ponding probably occurred in this area during deglaciation because of drainage obstruction by ice of the Athabasca Lobe. Moderately to well sorted gravels, constitute almost half of the sediments logged in the area. In the three sections immediately east of the Foothills (B10/R, B11/L and B12/L), a unit of 1 to 4 m of laminated silt and clay overlies the gravel units. This indicates the relative importance of the pattern of deglaciation in respect to the Athabasca and Brazeau Lobes. Unless the Athabasca Lobe continued to obstruct drainage to the east of the Foothills, it if difficult to explain the abundant evidence of waterlain deposits described above.

4.2.4 DRILLING LOGS:

Well logs and drilling logs from seismic shotholes were examined to determine the depth of the unconsolidated surficial deposits and to gain some insight into the composition of this material. Appendix 1 contains data obtained from the observations of stratigraphic characteristics of the deposits described in the drillers' reports. The locations of these sites are depicted on Figures 4.1 and A1.1. A summary of this material is as follows:

Locations 1 and 2 (Fig. 4.1) are within the western portion of the Foothills along structural ridges. At location 1 the total thickness of unconsolidated material was 12 m. The upper 0.6 m was described as clay but the drill site was investigated during this study and the surficial deposits were identified as a fine textured diamicton (till). Perhaps a local concentration of loess was present within the solum at the drill location. The drill log describes 3.4 m of clay and gravel underlying the upper unit. This unit is almost certainly a till. The lowest 8 m are described as gravel. From the topographic location of the drill site and the texture of the material examined near the drill hole, it is probable that these are outwash deposits or ice-contact sediments.

Location 2 was not found during this study but the cuttings from several seismic shotholes were observed in the immediate vicinity. The driller's report for this location does not specify the type of sediment which makes up the 7.5 m of unconsolidated material measured but this study suggests that it is fine textured diamicton. Bedrock at locations 1 and 2 is shale.

Within the central portion of the study area there is only one drill record and it comes from site 10 which is adjacent to Grave Flats. The driller's report describes 12.2 m of clay, sand and cobbles, overlying shale. The unconsolidated deposits were inspected during this study, and they are interpreted as diamicton, probably till.

Immediately east of the Foothills are four sites (8, 9, 11 and 12) which depict two different stratigraphic profiles. The two sample sites from the upland areas (11 and 12) contain very similar sequences of about 4 m of sand overlying (or coarsening downward into) 2 to 3 m of gravel. Sandstone bedrock underlies both of these locations. The other two locations (8 and 9) bracket, but are not within, the Brazeau River valley. The drill log from two profiles taken along the north band of the valley (8) displays a variable thickness (3 to 13 m) of till (clay and boulders) directly overlying bedrock whereas the southern location 9 shows a glaciofluvial or a glaciolacustine sequence of about 15 m of clay and sand overlying sandstone. In this study site 9 is described as lying within a pitted outwash train. This inconsistency of sediment description remains unresolved as the borehole from which this drill log was formulated was not found so the cuttings were not examined during this study.

Five drill logs are from the eastern portion of the study area (3, 4, 5, 6 and 7). All of the logs of these sample sites reflect a simple sequence of 1 to 14 m of diamicton directly overlying bedrock, except for sample location 4. This location is within the Uplands immediately east of the Blackstone River valley. The log describes 6 m of sand and gravel overlying bedrock. This description conflicts with the evidence gathered for this study and displayed on the Landforms Map. In this study the location in question was interpreted as ground moraine dissected by segments of meltwater channels. Either the description of the drill log failed to include a finer matrix which was observed during the fieldwork phase of this study, or the location drilled was within one of the meltwater channels. The latter explanation is the more plausible but the actual site was not found so

this could not be confirmed.

Problems exist in interpreting these data because the measurements are crude and the interpretations of the materials being drilled are often unreliable. This is especially true where bedrock consists of weathered sandstone or shale, as the material being drilled may be described as sand or clay instead of bedrock. Individuals from Alberta Research Council who have experience at interpreting drilling logs from the Foothills have cautioned that till deposits are frequently described as clay and because of the difficulty involved in drilling, the thickness of sand and gravel is commonly exaggerated. (D. Prosser pers. comm., 1978) Nevertheless, the logs do indicate that the glacial deposits generally form a relatively thin mantle over bedrock except for the area between the mountains and the confluence of the Southesk and Brazeau Rivers. In the valleys of the foothills bedrock lies, on the average, approximately 12 m below the surface. To the east of the foothills one well log indicates a thickness of surficial material as 13 m but most of the logs indicate bedrock at depths between 2 and 6 m. Most of the well logs indicate a sequence of till over gravel or sand with some logs indicating only glacial or glaciofluvial deposits. However, no instances were found of till overlying and underlying a nonglacial unit. It is therefore difficult to substantiate an argument for multiple glaciations in the area based on the limited, available, stratigraphic evidence.

4.2.5 TILL TRANSECT

Samples of the near-surface till below or within the soil *C* horizon were collected along a transect from the eastern slopes of the mountain front to the eastern limit of the study area (Fig. 4.1). Pebble counts were carried out, each of which incorporated a sample size of at least 150 pebbles. The results of this analysis are shown in Figures 4.5 and 4.6. Figure 4.5 expresses the lithologies in percentage terms whereas Figure 4.6 attempts to group the lithologies in terms of their source areas. The assumption made is that sandstone, siltstone and shale are derived largely from the Foothills, carbonates are from the Front Ranges and quartzite and chert are from the Main Ranges. This is simplistic in that sandstone, siltstone and shale are members of some of the rock units within the mountains but they do comprise a larger portion of the Upper Cretaceous and Tertiary Formations found in the Foothills. Figure 4.6 thus represents an adequate first

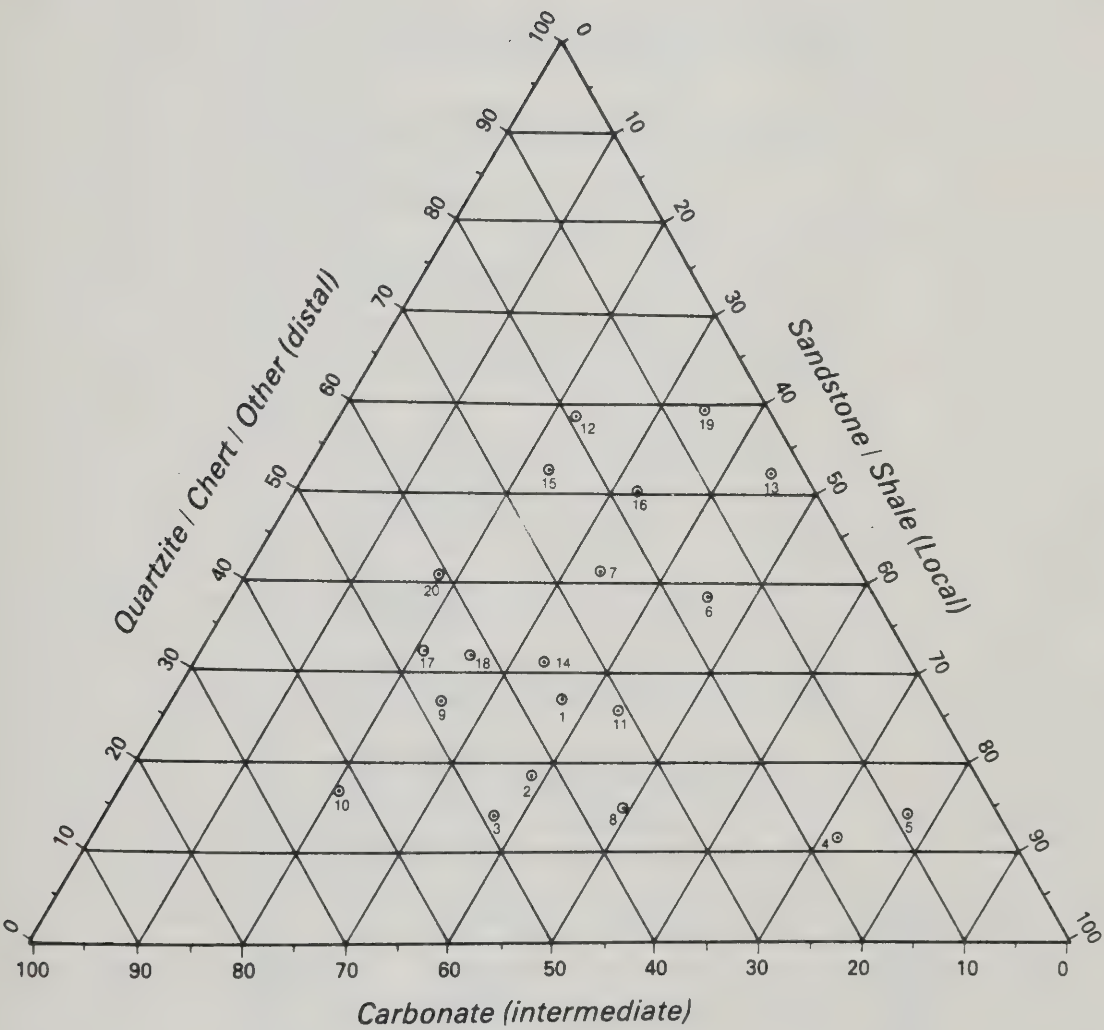

Fig 4.5 Relative Proportions of Dominant Lithologies, Till Transect Samples

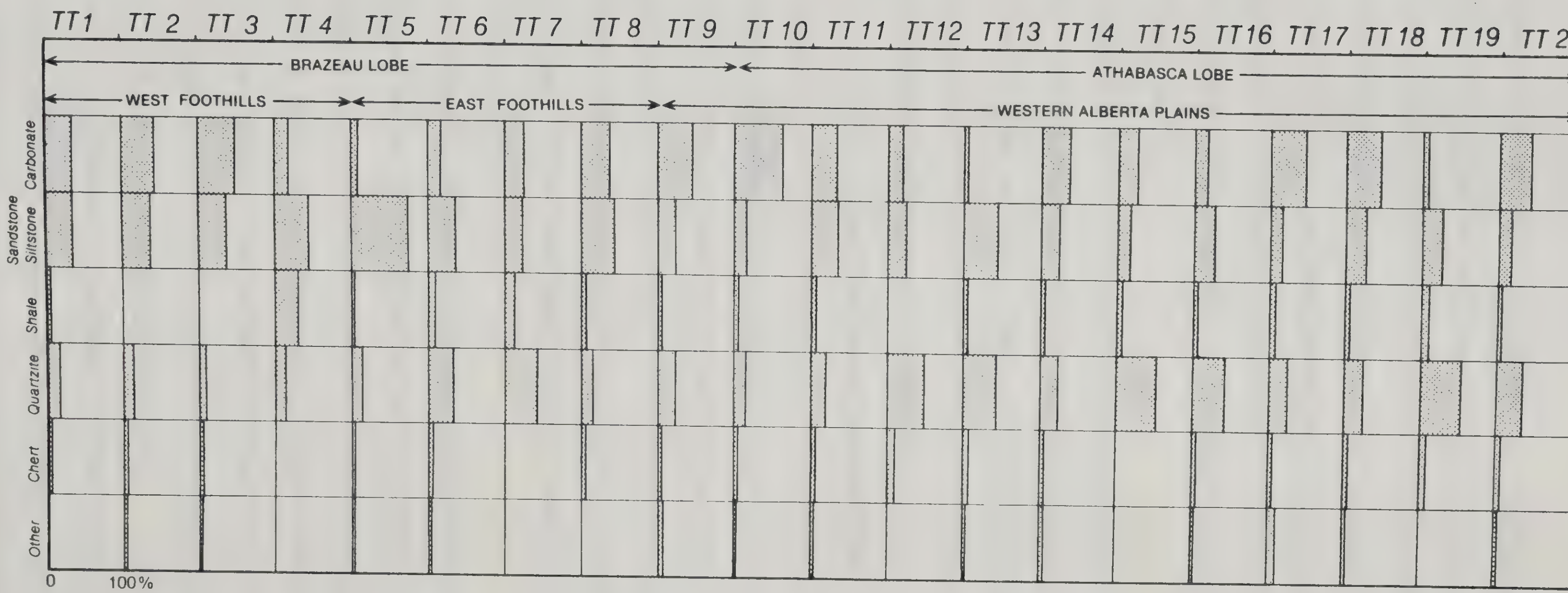

Fig. 4.6 Till Sample Lithologies

approximation of lithologic provenances. From Figure 4.6 it can be seen that no obvious trend of lithologies occurs across the study area. The samples from locations 4 and 5 appear to be anomalous. They were collected on the west and east sides, respectively, of the depression presently occupied by Muskiki Lake. The depression lies between, and transverse to, the Cardinal and Brazeau River valleys. Hence, it is reasonable to suppose that ice in this depression would have been substantially less active than the ice in the major valleys. On this supposition a larger proportion of local lithologies would be expected for locations 4 and 5.

Figure 4.6 indicates that the less resistant lithologies, such as sandstone, siltstone and shale, are more frequent in the tills of the Brazeau Lobe and the more resistant lithologies, such as quartzite, become more frequent in deposits of the Athabasca Lobe. A greater distance of transport for the latter is self evident. To assess the validity of this evidence, a number of tests were run using the MIDAS (Michigan Interactive Data Analysis System) program. The data were divided into sets of samples from two populations; one from the Foothills area, the other from the Western Alberta Plains. Seven lithologies were compared although carbonates, sandstones and quartzites account for 89 per cent of the total population. The standard deviations were, as Figure 4.5 implies, very high. For example, the average carbonate content was found to be 30 per cent with a standard deviation of 16 per cent. Similarly, for sandstones the mean value was 31 per cent and the standard deviation was 14 per cent. Quartzite had a mean of 27 per cent and a standard deviation of 15 per cent.

When the two sets of data were compared using a profile analysis it was found that the means of the percentage composition for each lithologic group were relatively closely clustered except for sandstone and quartzite. The mean percentage composition of sandstone was 41 per cent for the Foothills samples and only 25 per cent for the samples collected from the Western Alberta Plains. The quartzite statistics, on the other hand, show that the Foothills samples had a mean quartzite composition of 19 per cent whereas the samples from the Plains averaged 33 per cent.

A multivariate one-way analysis of variance was performed on the data to test for significant differences between the composition of the two sample populations. The calculated level of significance (0.017) does not support the hypothesis that there is a

significant difference in composition of till lithologies between the Foothills and that part of the Western Alberta Plains which is thought to have been overriden by ice of the Athabasca Lobe.

A point of potential interest for future investigations is the observed difference in the sandstone and quartzite compositions for the two areas mentioned above. Although the difference was not great enough to provide statistically significant results in this study, if a study was conducted which concentrated only on those two lithologies, and a much larger sample size, perhaps significant results could be obtained.

A distinctive pink-colored quartzite, believed to originate from the Gog Formation, is found throughout much of the study area. The presence or absence of such quartzite clasts may be used to distinguish between tills of the Brazeau and the Athabasca Lobes, because they appear to be much more common in material deposited by ice from the latter. Scattered clasts of pink quartzite have been found within gravels in the Foothills, but they are very infrequent and none was found in tills of this area.

4.2.6 SOIL DEVELOPMENT AND CALCIUM CARBONATE HORIZONS:

Soil development and pedogenic processes, such as the formation of clay minerals and the translocation of clays and water soluble salts, have received considerable attention from Quaternary geomorphologists. In fact, several books such as those by Morrison and Wright (1965), Pawluk (1969), Birkeland (1974) and Foscolos *et al.*, (1977), outline studies in which glacial surfaces of different ages are recognized partly on the basis of pedologic evidence. Karlstrom (1981) describes pedological evidence from southwestern Alberta which indicates that at least seven different ages of surfaces are expressed in the surficial geology of this area. Geosols are defined by Morrison (in Morrison and Wright, 1965) as soil-stratigraphic units which may be used as relative age dating units in the same way as rock-stratigraphic units. He recommends that geosols be dated and correlated strictly according to the rules of the Stratigraphic Code.

Birkeland (1974, p.263) outlined a number of problems which may be encountered when attempting to correlate or define soils on the basis of field evidence alone. He explained that, because of micro-environmental differences, there may be several pedologic facies for a given soil-stratigraphic unit. Nevertheless, he and others

have used criteria such as soil color, clay content and distribution in the solum, carbonate concentration and profile thickness as useful indicators of relative ages of surfaces. In fact the depth of the profile to the base of the B horizon has become a conventional criterion for describing the degree of development which the solum has undergone. Such investigators usually augment that evidence with data on the degree of weathering of surface clasts, landform preservation and stratigraphic position.

In the Rocky Mountains of Alberta a number of studies relating to soil development have been conducted. Studies by Beke (1969), Pettapiece (1970), Beke and Pawluk (1971), Knapik *et al.*,(1973), Brewster (1974), King and Brewster (1976, 1978), Howell (1977), Howell and Harris (1978) and Karlstrom (1981) deal with pedogenic development in this area. In the Yukon, Foscolos *et al.*, (1977) used pedologic evidence to interpret paleoclimates of the interglacial periods separating the glacial advances advocated by Bostock (1966). In their study, and in the work of Karlstrom (1981), the pedologic evidence augments the information which was originally interpreted from geomorphic evidence.

Reimchen and Bayrock (1977) suggested that the translocation of calcium carbonate may give valuable evidence on deposits of different ages in the Foothills. They identified three different ages of surfaces on this basis. Despite theoretical objections to the use of carbonate leaching as a relative-age dating technique, a test of the technique was conducted during this study. Thirty-six soil descriptions were recorded. At each location an attempt was made to choose a level, well-drained site with surface vegetation representative of the local area. On the basis of field observations most of the soils were classified as Brunisols, although some well-drained sites within the major valleys and on the Western Alberta Plains had soil profiles that would probably qualify as Luvisols.[3] A variable mantle of loess is found over much of the area. Up to 36 cm of

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[3] Subsequent to this study the Energy and Natural Resources Report #147 was published (Alberta E.N.R., 1980). This report is part of a series of physical land classification reports of the province. The western portion of the E.N.R. report covers the eastern portion of the area mapped in this study. In general they support the field observations presented in this study. They describe the soils of the area common to both studies as follows: "The soils of the system are mainly well to rapidly drained Eluviated Dystric Brunisols with a loamy sand surface horizon overlying gravelly sand parent material. These soils indicate strong leaching and have low pH values; usually 5.0 or less. Soils on till parent material have better profile development and include a clay accumulation (Bt) horizon. Orthic Gray Luvisols and Brunisolic Gray Luvisols are the major soils developed on till material."(Alberta E.N.R., 1980, p. 21)

loess were observed at some sites along the Western Alberta Plains. This silt-textured material significantly alters the texture of the upper solum. Hence, without extensive laboratory analysis it is difficult to determine the texture of the original parent material. This effectively precludes the use of translocation of fines in the solum as an indicator of relative age. It was therefore decided to use the total depth of the solum (**A** and **B** horizons) as a relative indicator of soil development. Birkeland (1974) and others have shown that generally the solum deepens and the clay content of the **B** horizon increases with age until a steady state is reached. If depth of the solum increases with time, and if the surficial deposits of the Foothills uplands and parts of the Western Alberta Plains were deposited by an earlier glacial event, one would expect to find deeper and better developed soil in these environments and less well-developed soil along the valley bottoms which were supposedly influenced by later ice advances (Reeves, 1973; Reimchen and Bayrock, 1977; and Rutter and Schweger, 1980). To test this hypothesis a plot of depth of solum and elevation was made. The results are depicted in Figure 4.7. From this graph any significant relationship between altitude and the solum thickness seems doubtful.

Another hypothesis tested using the soil data was that if Late Wisconsin ice was restricted to the major valleys, and only formed piedmont glaciers at the mountain front, older surfaces (pre-Wisconsin or Early Wisconsin) should be evidenced by deeper soil profiles at some distance along the valleys east of the mountain front. Figure 4.8 is a plot showing the depth of solum compared to the relative position of the site from the mountain front. The X-axis of the diagram is not to scale but the easternmost distal site is approximately 65 km from the mountain front. This graph illustrates that no significant difference exists between soils near the mountain front and those found on the plains immediately east of the Foothills.

Because the relative depth of calcium carbonate leaching was advocated by Reimchen and Bayrock (1977) as a potential dating device, therefore, at each soil measurement site the presence of precipitated calcium carbonate were recorded. Only visually observable carbonate rinds were noted as the high carbonate content of the parent material resulted in nearly all tills reacting effervescently when treated with dilute hydrochloric acid. It was found that immediately east of the mountain front abundant

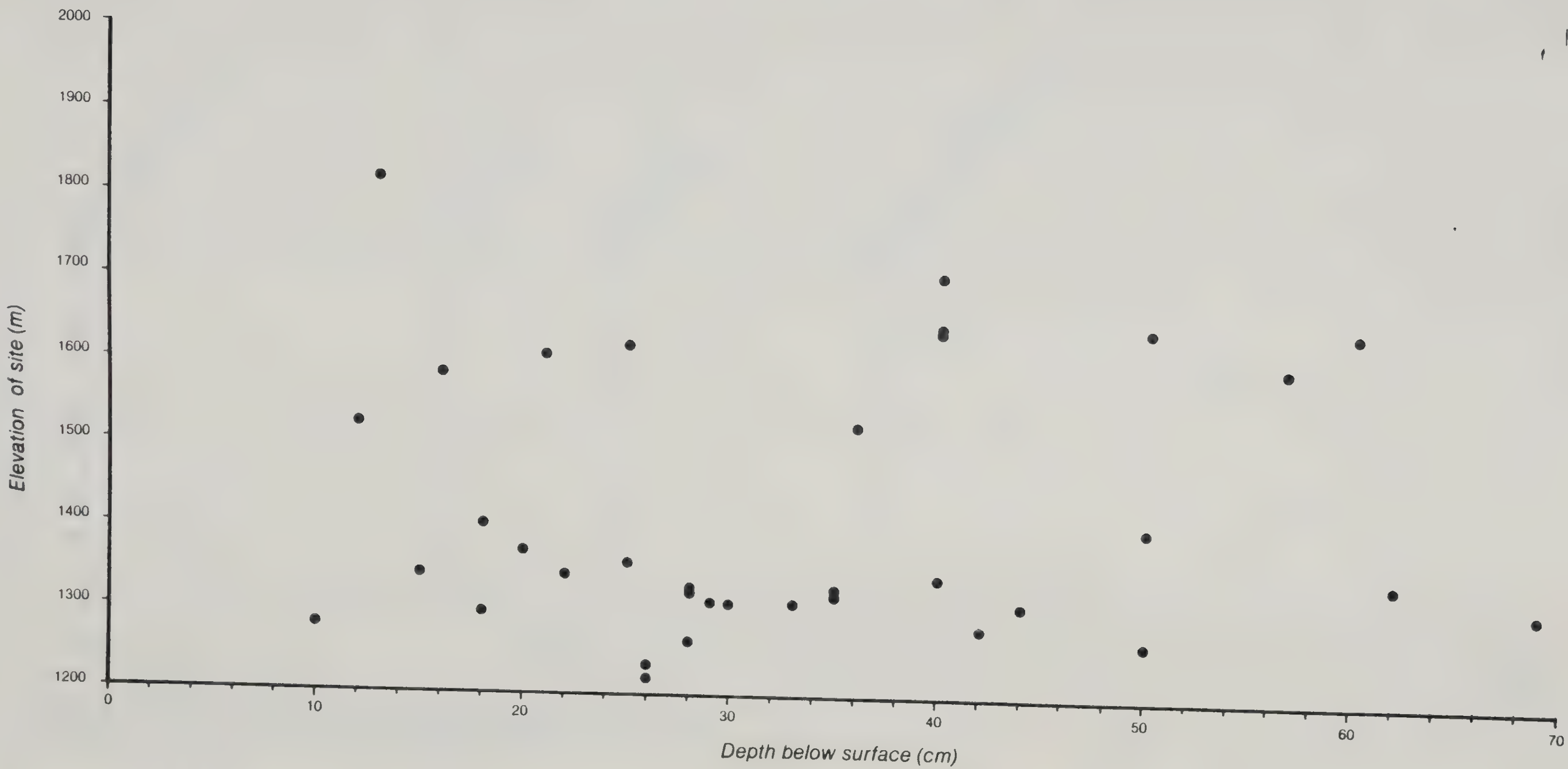

Fig. 4.7 The Relationship Between Solum Thickness and Site Elevation

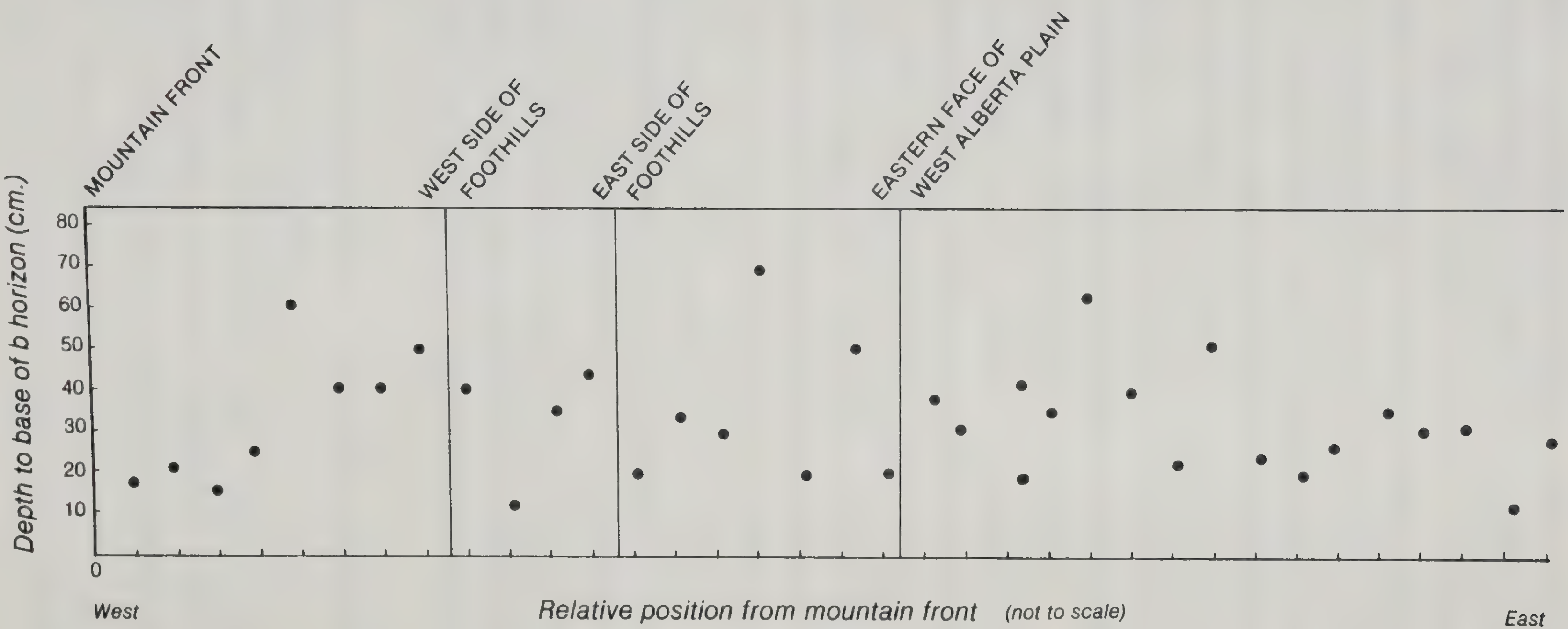

Fig. 4.8 Solum Depth Transect Across Study Area

carbonate precipitates exist from the surface to a depth of approximately 1.5 m. Frequently these took the form of a *Cca* horizon but in coarse textured parent material the precipitates were visible as rinds on the bottoms of surface clasts. It was not uncommon to find dripstone rinds up to 3 mm thick beneath surface clasts and in the *C* horizon rinds were observed on the upper surface of the clasts. The significance of this is not known, but it does indicate that carbonate has been mobile. Figure 4.5 indicates that carbonates are a major constituent of the tills. Birkeland (1974) illustrates how cool moist temperatures and slightly acidic soil solutions enhance the mobility of carbonates in the soil. It is therefore not surprising that in calcareous parent material the clasts in the upper 1.0 – 1.5 m of these deposits are usually encrusted with a carbonate rind and frequently some carbonate cementation of the finer fractions has occurred.

These observations do not, however, substantiate the carbonate leaching categories suggested by Reimchen and Bayrock (1977). They found that "young till" was leached 0' – 2' (0 – 0.6 m), "intermediate till" was leached 2'– 5' (0.6 – 1.5 m) and "old till" was leached to depths as great as 14' (4.3 m). Observations from the present study show that in the tills of the Western Alberta Plains little or no precipitated carbonate occurs in the upper 2 m of the deposits. The lower limit of leaching was not determined. A systematic study using the Chittick apparatus described by Dreimanis (1962) would possibly furnish additional information as to the actual extent of carbonate translocation but the fact remains that near the mountain front the tills cannot be significantly leached when travertine deposits occur on the bottom of surface clasts. With the variability of carbonate content in the parent material (illustrated in Figure 4.5), any simple causal relationship identified between the age of the deposit and the depth of leaching is likely to be spurious.

In summary, the major problems of interpreting pedologic evidence as indicators of relative landform age in the Foothills are:

(1) Local climates within the Foothills are highly variable. Aspect, temperature, soil moisture content, exposure and elevation change over short distances in this area. The local rates of pedogenesis will necessarily be influenced by these differences. Therefore, if soils of different ages are present within the Foothills the catinae would necessarily have a high degree of overlap and it is doubtful that any general

pattern could be identified with any confidence.

(2) The composition and texture of the parent material is not uniform throughout the region. The Landforms Map gives some impression of the variability of surficial deposits, but there are also significant variations that occur at a very local level. As discussed in Section 3.4.3 and in some of the stratigraphic descriptions, thin, discontinuous veneers of lacustrine or aeolian sediments are relatively common within the study area. It is quite evident that although these deposits are of secondary importance for geomorphic interpretations, they will have a major effect on pedogenesis.

(3) The area is heavily forested and there is obvious evidence of a long and complex fire history.

Birkeland (1974), King and Brewster (1978) and Howell and Harris (1978) have emphasized the effect that forest fires have on the pedologic processes. Birkeland (pers. comm. 1976) had serious doubts about the applicability of any pedogenic criterion for relative dating purposes in areas that have been intermittently exposed to fire.

4.3 LAKE AND BOG CORES

The radiocarbon dates obtained from the cores of sediments extracted from lakes and bogs in the study area are listed in Table 4.1. The bog dates indicate that peat began to accumulate in this portion of the Foothills approximately three to five thousand years ago. These findings agree with basal dates obtained from other bogs in west-central Alberta by a number of researchers from the University of Alberta. (T. Habgood and C. Schweger, pers. comm., 1979). With the exception of a date of approximately 6240 years B.P. (Muskrat Bog, Habgood and Schweger, pers. comm.), all of the available bog dates range between 3035 and 4735 years B.P. This may indicate that the post-Altithermal climate of west-central Alberta became cooler and more moist approximately 4000 years ago and has remained so since then.

CORE LOCATION	CORE LENGTH	SAMPLE	AGE (yrs. B.P.)
Grave Flats Bog	2.08 m	S-1525	4735+/-110
Thunder Lake Bog	1.6 m	S-1524	3035+/-85
Muskiki Lake	3.43 m	S-1523	8300+/-435
Fairfax Lake	5.5 m	S-1630	8970+/-90
Fairfax Lake	10.15 m	S-1705	11,225+/-120

Table 4.1 Radiometric Ages of cores obtained from Study Area.

4.3.1 MUSKIKI LAKE CORE

Muskiki Lake occupies a structural valley between the Brazeau and Cardinal Rivers. The elevation of this lake is 1525 m (see Landforms map). The core extracted from Muskiki Lake was 3.43 m long and was primarily composed of silt and organic sediments, although near the base of the core the texture became much coarser. Samples were taken at 5 cm intervals along the length of the core. Tests with dilute hydrochloric acid indicated that nearly all of these samples had a significant carbonate content. A band of volcanic ash, believed to be of Mazama origin, was observed between 68 and 75 cm from the top of the core. Gastropod and bivalve mollusc shells were common throughout the core. At 38 cm, 48 cm, 55 cm, and 69 cm from the top, the shells and shell fragments formed a major constituent of the sediments. The incidence of shells decreased with depth and below 162 cm the shells were infrequent or absent.

Subfossil plant fragments existed throughout the core but in the upper half of the core they comprised a fine-textured gyttja whereas toward the base of the core macro-fragments of well-preserved mosses and vascular plants were common.

The Muskiki Lake core may not have reached the most basal sediments because a definite bedrock contact could not be detected with the coring tool or by probing. The coarser texture and more compact nature of the material at the base of the core suggests that the underlying material was deposited in a very different environment than that of the sampled sediments. In addition, 16 fragments of subfossil bryophytes (*Drepanocladus fluitans*) were extracted from the sediments of the lowest 5 cm portion of the core. Janssens (1978, p. 4) identified the sub-fossil plant material and suggested that the habitat characteristics of the species are as follows:

> "*Drepanocladus fluitans* grows in mires, pools and lakes. Submerged forms are quite distinct but undercollected. The species is non- calcareous and is found in poor fens (pH values of 3.6 and 4.2 in Stordalen) (Martensson, 1956). It indicates a highly dynamic environment (changes in water level and/or anthropogenic influences), (Landwehr 1966). It occurs from the lowlands up into the low-alpine region of the mountains (Nyholm 1965)."

The intolerance that this species shows for calcareous environments and deep stable lakes suggests that an early postglacial environment with a plant community comprised of aquatic and hygrophytic vegetation and rapid sedimentation rates is most plausible. The preservation of the plant fragments is sufficiently good that leaves were found attached to stems. This rules out the possibility of long-distance transport or re-working of these fragments. The lake presently collects only local runoff with no permanent streams feeding into it. A single outlet stream drains into the Cardinal River. Sedimentation rates are therefore low and it is unlikely that rapid rates of sedimentation have occurred since the area was stabilized by vegetation.

The evidence from this core is difficult to interpret because the radiocarbon age (8300 +/- 435 years B.P., S-1523) of the plant remains extracted from the basal sediments is young and not likely to represent an early postglacial date. Perhaps the radiocarbon date is inaccurate as the sample size was very small. The reason for this is that only plant macrofossils were dated to reduce the chance of carbonate

contamination. The large sigma value (+/- 435 years) tends to erode confidence in the accuracy of the date. Even if the sigma value is added to the date, the 8700 year B.P. date thus obtained is considerably younger than other early postglacial dates obtained for the general region. For example, Ms. E. Bombin, (Department of Botany, University of Alberta, pers. comm., 1980) obtained two radiocarbon dates from Mary Gregg Lake which is located 42 km northwest of Muskiki Lake. The basal date for the 7 m core obtained from Mary Gregg Lake was 10,400 +/- 220 years B.P.(GSC-2935) and a shorter core (just over 4 m) obtained from another part of the same lake produced a basal date of 11,000 +/- 120 years B.P.(GSC-2997).

Bombin (1982) feels that these dates are probably inaccurate as coal fragments and carbonate was observed in the sediments. Perhaps her concern is unjustified in that if 'dead' carbon was accidentally included in the material sampled, the dates should be too old rather than too young. Because the C14 decay curve is logarithmic the older the sample the less significant a small amount of 'dead' carbon will be. M. Striver of the United States Geological Survey, Denver (pers. comm., 1980) considers that contamination of 1% could essentially be ignored for 20,000 year-old dates and 5% contamination would bias, but not invalidate the date. Most of the lakes which have been cored and dated in Alberta reflect either early Holocene (10,000 yrs. B.P.) basal dates or dates of approximately 4,000 to 6,000 yrs. B.P. which probably relates to Altithermal desiccation (Schweger, *et. al.*, 1981). If the Mary Gregg sediments are contaminated they would probably reflect one of the two time intervals discussed above or Mary Gregg lake was formed by some unknown event(s) between these two periods. The lake lies within a structural valley with no evidence of drainage obstruction due to landslides or recent tectonic activity, therefore it is safe to assume that the local water balance was instrumental in filling the lake. The unresolved question is then whether or not the date is sufficiently accurate to record approximately deglaciation in the area or if contamination has contaminated a Holocene date sufficiently to reflect a late-Pleistocene age.

The core from Fairfax Lake (Table 4.1) furnished a basal date of 11,225 +/- 120 years B.P. (S-1705). Obviously the Muskiki Lake date of 8300 +/- 435 years B.P. (S-1523) does not agree well with the date from Fairfax Lake and those from adjacent areas. In the opinion of the author the 8300 date is questionable and additional samples

are required from this lake to substantiate or reject it. Although the date is questionable, the sedimentary environment inferred from the bryophyte evidence suggests that the core probably extends into the sediments deposited under very different environmental conditions. A very plausible explanation for the rapid sedimentation rate and the decrease in pH would be that the lake basin was being filled from an adjacent glacier. This interpretation implies that it is very unlikely that the area was unglaciated throughout the late Wisconsin glaciation as Reimchen and Bayrock (1977) and others (discussed in Chapter 6) have suggested.

1.3.2 FAIRFAX LAKE CORES:

Two cores were extracted from Fairfax Lake. A 5.5 m core was obtained in January, 1979, and the basal sediments furnished a date of 8970 +/- 90 years B.P (S-1630). At the time of the core extraction it was felt that perhaps this core did not extend to the bottom of the sediments. After the radiometric date confirmed that suspicion, it was decided to extract another core. In March of the same year a 10.15 m core was obtained and this furnished a basal date of 11,225 +/- 125 years B.P.(S-1705).

Fairfax Lake occupies an erosional valley which was probably created by glacial meltwater. The ends of the valley were obstructed by younger deposits. This valley transects the easternmost ridge of the Foothills. The present elevation of the lake is 1316 m and the coring was conducted through 5.8 m of ice and water. The lake is sustained exclusively by a local catchment area and drains to the northwest into the Pembina River.

The Fairfax Lake core, like the Muskiki Lake core, was largely composed of gyttja in the upper portions. The matrix material was silt and clay with no macrofossils. A 0.5 cm thick layer of volcanic ash was identified at 285 cm from the top and an attempt was made to concentrate sufficient shards for electron-microprobe analysis using the technique described by Smith and Westgate (1969). No shards could be separated by either the author or Dr. Smith. The lack of shards may indicate that the ash has been reworked (D.G.W Smith, pers. comm.) A second ash band approximately 6 cm thick was identified between 470 and 477 cm. This ash was separated and the shards were prepared and analyzed by the author under Dr. Smith's supervision. The electron

prepared and analyzed by the author under Dr. Smith's supervision. The electron microprobe analysis (Table 4.2) confirmed this tephra to be Mazama Ash, approximately 6,600 years old.

Below the Mazama Ash the incidence of fine textured organic material decreased, although stem and seed macrofossils were in evidence. No coal fragments were observed above 800 cm. Coal was found below this level but it was only below 970 cm that coal-rich bands and larger, visible, coal fragments were concentrated.

A 10 cm slug of the core from 744 and 754 cm below the top was removed and used for the radiometric date. It was felt that the transition from sand and silt, with coal inclusions, to fine silt represented a change in the depositional environment from a glaciolacustrine to lacustrine water body. The coal fragments contained in the sediment are highly significant. According to field observations and the geological maps of the area (MacKay, 1940), there are no significant coal outcrops in the Fairfax Lake basin, but rather shale and sandstone facies of the Brazeau Formation predominate. In the valley immediately west of the ridge that Fairfax Lake lies within, coal outcrops of the Edmonton Formation are common. They may also be observed along the banks of the Pembina River and a thick seam of coal is currently being strip mined along the Lovett River 16 km to the northwest of the lake.

Obviously, the coal has been transported prior to deposition in the lake sediments. Although it may have been transported from within the basin, the evidence previously cited indicates that this was unlikely. Alternatively, it might possibly have been washed from the till which mantles a large part of the basin and redeposited in the lake. This is not probable because comminuted coal tends to weather relatively rapidly, as can be documented at the abandoned collieries within the study area. There are no streams flowing into the lake so re-deposited fluvial sediments cannot be the explanation for their existence. The most plausible explanation is that they were transported in by glacial meltwater issuing from the glacier which is thought to have occupied the valley to the west of the lake, and the abrupt decrease in coal inclusions in the lake core is thought to reflect the transition from a proglacial lake to a nonglacial lake environment. The sedimentary evidence substantiates this interpretation in that the deeper sediments are coarse textured and they grade upward into silty clay, then clay-gyttja sediments. The

	Average composition of Fairfax Lake Samples 1&2	Average composition of Mazama ash (Smith, D.G.W., Pers. Comm.)
Na_2O	5.25	5.15
MgO	0.56	0.54
Al_2O_3	14.47	14.42
SiO_2	72.53	72.59
Cl	0.20	0.18
K_2O	2.79	2.70
CaO	1.75	1.71
TiO_2	0.56	0.48
FeO	1.94	2.08

Table 4.2 Chemical composition of tephra shards from Fairfax Lake core as determined by electron microprobe analysis (wt. %)

most abrupt transition in texture occurs just below the area of the core that was sampled for radiometric dating.

One of the premises upon which pollen analysis is predicated is that the pollen preserved in the sediments accurately reflects the local vegetation at the time of deposition. There are obviously a number of weaknesses of this assumption. Exotic pollen may be transported into the area by wind, ice or water; some pollen types do not preserve well in lacustrine environments (e.g. *Populus*) and the local pollen dispersal from the local catchment area may not be accurately reflected in the sediment. Nevertheless, it remains a useful tool with which to reconstruct paleoenvironments. One conclusion that one may draw from pollen analysis can reach with reasonable certainty is that the absence of pollen of a given species (assuming that the species usually contributes to the local pollen rain) implies that the species in question was not likely present in the local area. Palynological interpretations are therefore frequently inferences drawn from negative evidence: a somewhat unusual scientific procedure. An additional complication for palynological studies pertaining to late-Pleistocene/early-Holocene events is that the absence of pollen from a specific species may reflect the paucity of propagates in the area of study which may be misinterpreted as environmental limiting factors. The final note of caution regarding the interpretation of pollen spectra is that some species of

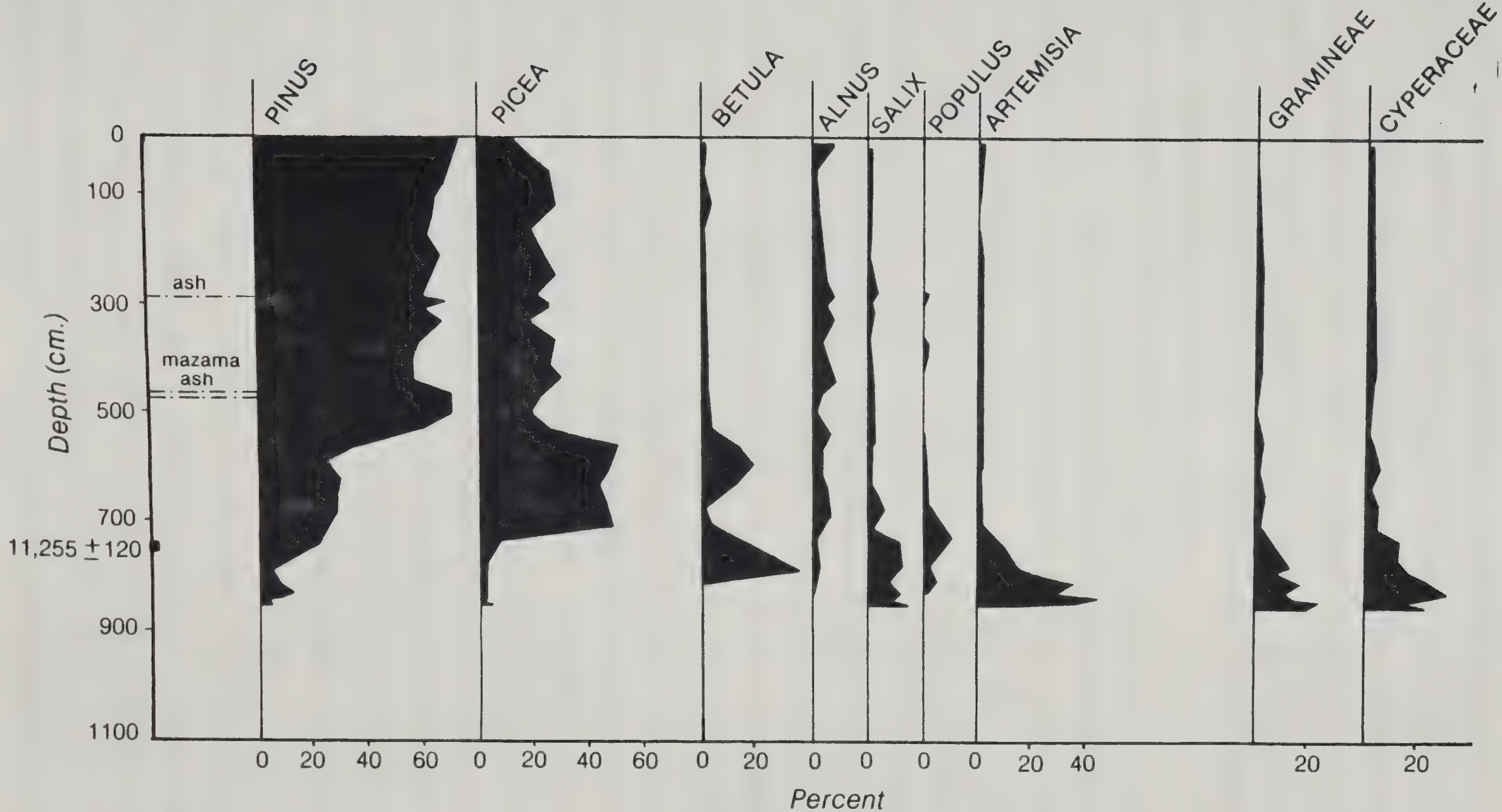

Fig 4.9 Pollen Diagram: Fairfax Lake Core

Analyst: T. Habgood

plants exist in diverse forms but the pollen grains are indistinguishable or at least easily confused. Birch (*Betulaceae*) is a good example. The pollen from dwarf birch (*Betula glandulosa*), paper birch (*Betula papyrifera*) and water birch (*Betula occidentalis*) are all included in one category on the pollen diagram (see Figure 4.9) but the habitats conducive to colonization by the different species of birch range from mesic-subarctic, through mesic cool-temperate to hygric-temperate. Some ambiguity and thus some conflicting interpretation of pollen spectra are therefore unavoidable.

The interpretation of the pollen data obtained from this core is controversial. Schweger (*et al., 1981*) suggest that the pollen evidence indicates an early treeless shrub tundra environment during the time when the sediments below the 800 cm level were deposited. The high percentages of shrubs (*Betula, Salix, Populus and Artemisia*) as well as the relatively large amounts of pollen from the grass and sedge families (*Gramineae* and *Cyperaceae*) are cited as support for this interpretation. Schweger (pers. comm. 1979) interprets the minor arboreal component in the lower portions of the core as exotic, wind-borne pollen from forested areas to the south. Schweger *et. al.*, (1981) have interpreted the lower portion of the Fairfax core to reflect the conditions during the late-Wisconsin maximum. However, he does state that such an assemblage may represent a pioneer plant community that became established shortly after ice recession (Schweger *et. al.*, 1981, p. 50). It is this alternative interpretation that is favored by the present author. The lower portions of the Fairfax core (below 850 cm) are almost completely barren of pollen. Counts of less than 100 pollen grains per microscope slide were normal for the sediment below 850 cm (Habgood, T. pers. comm., and personal observations). Very low pollen counts may reflect relatively rapid rates of sedimentation, relatively low amounts of airborne pollen, or a combination of both. Modern pollen rain studies for high latitude environments such as Baffin Island indicate that 300 counts per microscope slide are normal and only occasionally are counts as low as 100 observed (Andrews and Nichols, 1981, p. 388). If Fairfax Lake existed throughout the late-Wisconsin advance it would have probably had a very low sedimentation rate because it is sustained only by local catchment. Loess would have probably provided most of the mineral sediment input. Almost certainly the local catchment area would have been fully vegetated with at least a tundra biome and probably a significant amount of

exotic pollen would have been transported in by wind from the south. This scenario is not supported by the paucity of pollen, nor by the coarser texture of the sediment in the lower portion of the core. A more plausible explanation is that the lower portions of the core reflect more rapid sedimentation and sparse surface vegetation during deglaciation. The recent pollen evidence from northwestern Canada presented in Ritchie (1984) tends to support this interpretation. Most of the pollen diagrams which contain evidence from earlier than 10,000 yrs. B.P. show a high percentage of birch *(Betula)*, willow *(Salix)* and grass and sedge *(Gramineae)*. Lateral Pond, located in the South Richardson Mountains, was cored specifically because it is a moraine-dammed lake which was probably formed by the ultimate advance of the Laurentide glacier. Even in this environment, which must have been very severe, the total pollen influx in the sediments which are between 16,000 and 12,000 years old was approximately 1,000 - 3,000 grains per square centimetre per year (Ritchie, 1984, p. 104). These values are low but still adequate to allow the construction of a meaningful pollen diagram (Ritchie, 1984, p. 105) which differs substantially from the one presented in this study (Fig. 4.9).

The lowest portion of the Fairfax Lake core with significant amounts of pollen (about 850 cm below the top) contains pollen from an association of herbs, shrubs, grasses and sedges; exactly the type of plant community that one would expect during primary colonization of a recently exposed surface. The increased presence of willow and poplar (*Salix* and *Populus*) between the depths of 700 and 800 cm in the core supports this interpretation as both of these species are rapid colonizers of disturbed sites because of their abilities to reproduce vegetatively as well as from wind-borne seed dispersal. The poplar count is probably underrepresented in Figure 4.9 because of the relatively fragile nature of its pollen grains. The sharp spike in the birch (*Betula*) profile at this level may reflect arboreal birch (e.g. *Betula Papyrifera*) colonization augmenting the shrub birch component of the pollen rain.

At approximately the 700 cm level, the coniferous arboreal component begins to dominate the pollen spectra. Spruce (*Picea*) appears to have been the dominant pollen contributor at about 11,000 years B.P. but at approximately 8,000 years B.P. (extrapolated from the dated level and the Mazama ash band) pine (*Pinus*) replaced spruce as the dominant pollen component. This change may reflect the vegetational change

induced by the warmer conditions associated with the Altithermal. Weak support is provided for the hypothesis by the slight decline in sedge (*Cyperaceae*) and the slight increase in grasses (*Gramineae*) which were synchronous with the sudden increase in the pine component. The decrease in birch and poplar during this time interval would seem reasonable if this interpretation is correct but willow pollen increases slightly instead of decreasing as one might expect.

It is unlikely that further interpretation of this, or subsequent cores, will conclusively resolve the conflict between the two paleoenvironmental interpretations. Nevertheless, it would seem to be much easier to explain the paucity of pollen in the lower portion of the core by the arguments expressed in this study than to try to explain how a lake in a paraglacial environment could fail to record significant amounts of pollen when plants would have doubtlessly existed at least within the catchment basin. Ritchie's (1984) pollen diagrams from the non-glaciated portions of northwestern Canada clearly illustrate that significant pollen rains were being deposited in those areas throughout the last glacial event.

An additional problem confronting the interpretation of the Fairfax Lake core and other sediment cores from the Foothills of southwestern Alberta is the absence of Glacier Peak tephra in the sediment column. Porter (1978) describes evidence of a series of volcanic eruptions which have been traced to the North Cascade Range in northern Washington which is much closer to the study area than the sources of the Mazama and St. Helen's tephras which are commonly found in lake sediments in Alberta. Porter (1978) suggests that at least three and perhaps as many as nine significant eruptions occurred from the Glacier Peak volcano between 12,750 and 11,250 years ago. Only at Manyberries in southeastern Alberta has the tephra been identified. Perhaps the ash falls from these eruptions were transported away from this area by winds (this is almost certainly the case for event M which appears to have travelled nearly due south) but the pattern of the tephra plume from event G should have extended well into southwestern Alberta.

5. ESTIMATED EXTENT OF GLACIATION

5.1 INTRODUCTION

As a result of problems associated with time, space and scale, a major goal of research in the earth sciences is to organize the observational data into a comprehensive explanatory model. Once generated, the model must be tested. Traditionally the three types of tests include verification, falsification and confirmation of the model. If the explanation is verified by several lines of evidence and it is not falsified by incompatible data, it is usually provisionally accepted and confirmation becomes an on-going process which may take a long period of time and numerous additional studies. Frequently, attempts to confirm the hypotheses result in minor adjustments or qualifications of the model in order to accommodate the new evidence. Occasionally the rejection of one (or more) hypothesis results in the rejection of the model.

In natural science, when dealing with past or future events, it is impossible to conclusively verify explanatory models. It is therefore necessary to concentrate on the least ambiguous evidence and to place considerable importance on the confirmation of the hypotheses. At any time new evidence may be produced which will falsify the explanation. Some explanatory models are only falsifiable in principle as there may be physical or temporal constraints which prevent the completion of related investigations. Nevertheless, the type of evidence which might falsify the model must be clearly stated or implied.

This chapter summarizes observations which are considered to be meaningful for the proposed model that ice from the Brazeau River valley merged with a glacier from the Athabasca Valley along the eastern slopes of the Foothills during the last major glaciation. It also discusses a set of observations which might falsify or confirm that model.

5.2 ESTIMATES OF ICE LIMITS AND THE EXTENT OF GLACIAL EROSION

THEORETICAL ESTIMATES

Flint (1971) calculated that during the Late Wisconsin maximum the snowline was lowered by approximately 1,000 m at mid-latitude continental locations. This estimate appears to have been accepted or confirmed by subsequent work (Sugden and John, 1976; Pierce, 1979; and various CLIMAP publications). Barry (1973) estimates that within the Rocky Mountains of Montana and Wyoming the snowline was probably lowered by at least 800 m and probably as much as 950 to 1200 m below the present elevation. Mears (1981) suggested a decrease in mean annual temperature of 10 to 15 degrees Celsius which would lower the snowline by 1000 to 1500 m (assuming the dry adiabatic lapse rate).

Immediately west of the study area, contemporary glaciers occupy the headwaters of most of the major river systems and permanent snow caps are present on most of the higher mountain peaks. The present snowline varies in response to a number of micro-climatic variables such as aspect, orientation with respect to the prevailing wind, and angle of slope, but an elevation of approximately 3000 m is perhaps an acceptable crude estimate of the present snowline for areas which are not in close proximity to existing icefields or glaciers. Published information from the Inland Waters Directorate (1978) indicates that the snow line on the Athabasca Glacier is as low as 2300 m. If a conservative estimate of the Late Wisconsin snowline as being approximately 1000 m below present is assumed, it can be demonstrated that most of the First Range, and the higher outliers such as Chungo Ridge and Red Cap Mountain, would be well above the snowline and therefore part of the accumulation area for local glaciers. A review of Figure 1.4 illustrates this point. A decrease in summer temperature of only 6 degrees Celcius from a general climatic deterioration, katabatic influences, or both, would lower the snowline to near the present community of Luscar and only the lower portion of the Athabasca River valley and the area to the east of the Foothills would have a mean summer temperature of greater than zero degrees. A deterioration of 10 degrees would move the snowline north and east to include all of the Foothills and

only the extreme northeastern corner of this study area would be below the snowline during the summer months. In either case, with such a large area above the snowline it becomes difficult to explain how small valley glaciers within the major mountain valleys could remove all of the accumulated ice unless, concurrent with the deterioration of temperature, there was a decrease in precipitation. The paleoclimatic reconstructions discussed by Barry (1973) indicate that the precipitation amounts on the eastern slopes of the Rocky Mountains of Montana, Wyoming and Colorado probably did not change significantly during the Late Wisconsin/Late Pinedale event. It is therefore most plausible that major valley glaciers formed within, and issued from, the Rocky Mountain valleys to the west of the study area.

These crude calculations imply that the proponents of the minimum-extent model of Late Wisconsin glaciation, as discussed in Denton and Hughes (1981), would have to postulate a reduced precipitation regime or intensive sublimation from katabatic winds in order to explain the glacier limits because the temperature regime must have been periglacial.

OBSERVATIONS AND PROJECTIONS

The maximum elevations to which former glacier ice surfaces reached may be approximately determined by identifying the upper limits of glacial erosion and deposition. These are probably conservative indicators of the actual ice limits, but the errors should be relatively small and consistent within a local region. The geomorphic evidence upon which the glaciation limits were based for this study includes ice-smoothed valley walls, breaks-of-slope, rounded cols, rounded ridge crests and summits, relict meltwater channels, erratics and lateral moraines.

The glaciation limits near the headwaters of the Brazeau River system, in valleys of both the south and the north forks, were found to be at elevations of about 2255 metres. In the eastern Front Ranges, and along the eastern face of the mountains, the limit declines to an elevation of approximately 2195 m. This limit was consistent throughout the Job Creek Valley and the Cairn, Southesk, and Brazeau River valleys as well.

To the east of the Front Ranges, along the Bighorn Range, rounded summits and subdued ridges exist below an elevation of 2195 m whereas steep slopes and rugged

topography exist above this level. Similar evidence of former ice limits was observed on Red Cap Mountain and limestone erratics were found near the Grave Flats Lookout Tower at an elevation of about 2075 m. East of the Bighorn Range quartzite erratics were found near the summit of a 2075 m high ridge. Therefore, surfaces at elevations above approximately 2195 m in much of the study area were probably ice-free during the glacial maximum (see Figure 5.1). It has been argued (Reimchen and Bayrock, 1977) that this glacial limit reflects a pre-Late Wisconsin advance. No radiometric dates are available from deposits associated with the upper glacial limit. However, no convincing geomorphic evidence was found of a hiatus in the landform suite such as breaks-of-slope, moraines or better developed soils which would indicate multiple events such as those described for the Waterton - Castle River area by Stalker and Harrison (1977) and Karlstrom (1981). The existence of erratics at relatively high elevations tends to support the suggestion that the upper ice limits probably relate to a Late Wisconsin advance, for the following reasons. The recessive nature of local bedrock, combined with steep slopes and a moist cold-temperate climate, tend to promote rapid rates of mass movement. An unpublished study of solifuction in the Kananaskis Park has documented rates of surface transport in the order of 0.5 cm per year (Smith, D. in prep.). Smith's study also illustrates that 10 metres of movement have occurred in solifluction lobes over the last 4000 years in his study area. Solifluction lobes and turf-banked terraces on some slopes above the timberline indicate that the surficial material is undergoing significant mass movement in this study area as well. In any case, it is apparent that erratics in the surficial deposits would likely have been transported a considerable distance down-slope or buried over a period of a few tens of thousands of years. The preservation of erratics under these conditions makes it fairly improbable that they were deposited during a pre-Late Wisconsin advance. This is particularly evident when one considers that if they had been deposited prior to the last glaciation they would have been exposed to a periglacial environment with accelerated rates of near-surface mass movement. In fact, the most probable explanation for the paucity of erratics at higher elevations is the accelerated rates of mass movement in these areas.

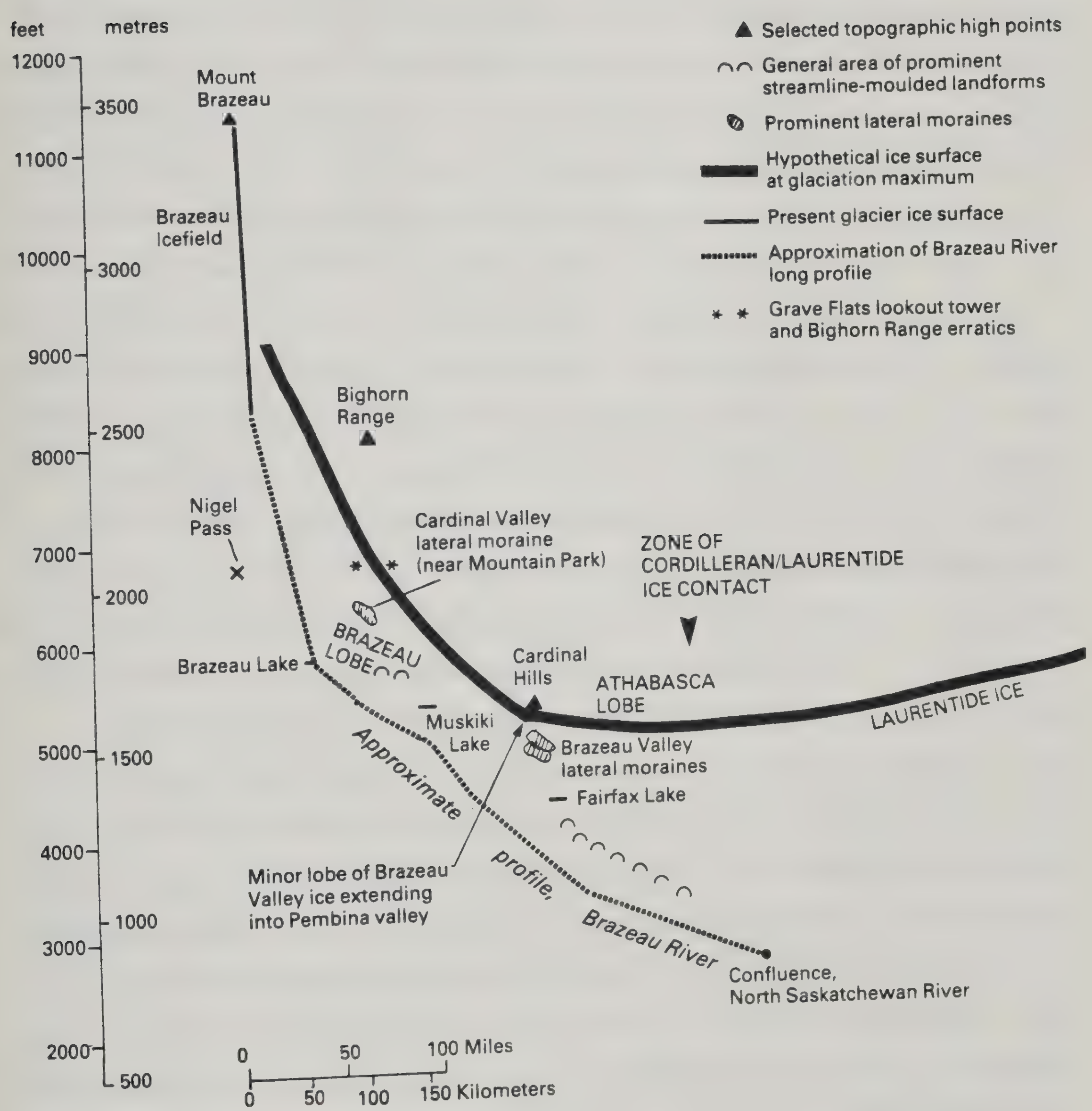

Fig. 5.1 Longitudinal profile of ice surface at glacial maximum

The source of the ice which occupied the Brazeau River valley is an important consideration. Haynes (1972) demonstrated a correlation of approximately 0.71 between the cross-sectional areas of glacial troughs and the sizes of contributing drainage basins. Earlier, Penck (1905, cited in Cotton, 1958) recognized that the cross-sectional area of a glaciated valley is proportional to the amount of ice that flowed through it. A crude analysis of the relative contributions of tributary valleys to the Brazeau Glacier system was obtained by comparing the cross-sectional areas of portions of the valleys below the observed upper limits of glacial erosion. No corrections were made for cross-sectional curvature of the ice surface or postglacial infilling of the valleys. It should be noted that the upper ice limits defined for Nigel Pass, (see Figure 5.1) and the unnamed pass immediately to the north of it, were about 2255 m. This is 30 m lower than that recorded by Roed *et al.*, (1967, p.631) for the Yellowhead Pass area and 60 m lower than the upper ice limit observed by McPherson (1970, p.12) on Mount Murchison. These other recognized ice limits are from areas northwest and southwest of Nigel Pass and are therefore approximately consistent with the ice limits identified in the present study. The results of this investigation are depicted in Figures 5.2 – 5.5.

Nigel Pass, and the unnamed pass immediately north of it, are the major corridors into the Brazeau River valley from the Main Ranges to the west. The valleys of the Cardinal River, Ruby Creek, the Southesk River and Job Creek are drainage basins which lie entirely within the Front Ranges. The cross-sections of these valleys were measured immediately upvalley of their intersection with the Brazeau River valley or at the point of exit from the First Range. Even at this crude level of comparison it is obvious that the glaciers which occupied the valleys within the First Range were much larger, and therefore discharged proportionately more ice, than ice streams which crossed the passes along the divide separating the Front Ranges from the Main Ranges (see Figure 5.4). This is, to some degree, different from the situation in the North Saskatchewan and Athabasca valleys where major intermontane valleys appear to have contributed large amounts of ice from west of the Front Ranges. Figures 5.3 and 5.5 illustrate that despite the relatively small size of the cols between the Main Ranges and the Front Ranges, the cross-sectional area of the Brazeau River valley at the point of exit from the First Range is more than one-half the size of the North Saskatchewan River valley at the analogous

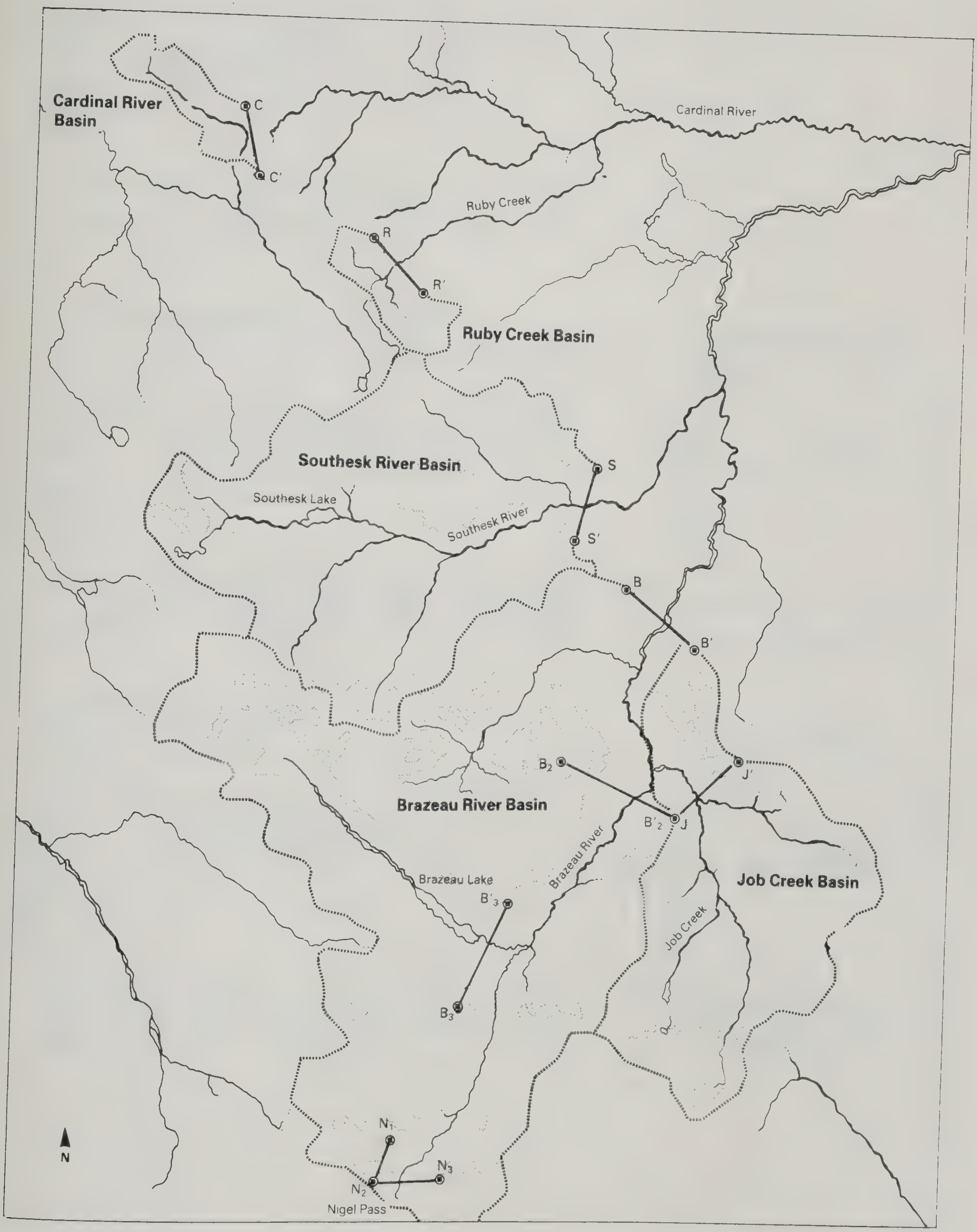

Fig. 5.2 Locations of valley cross-sections and drainage basins

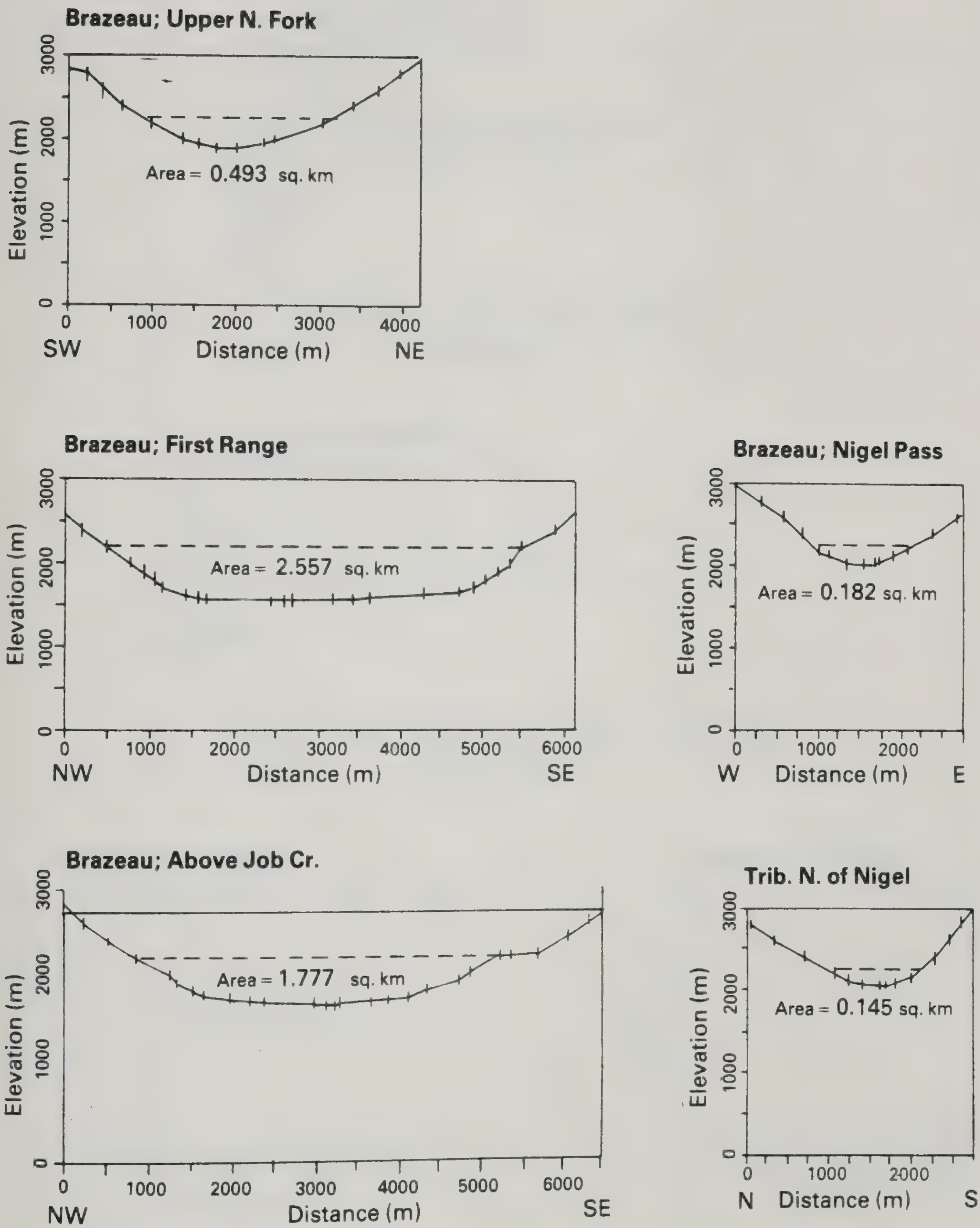

Fig. 5.3 Cross-sections of valleys within the Brazeau drainage basin

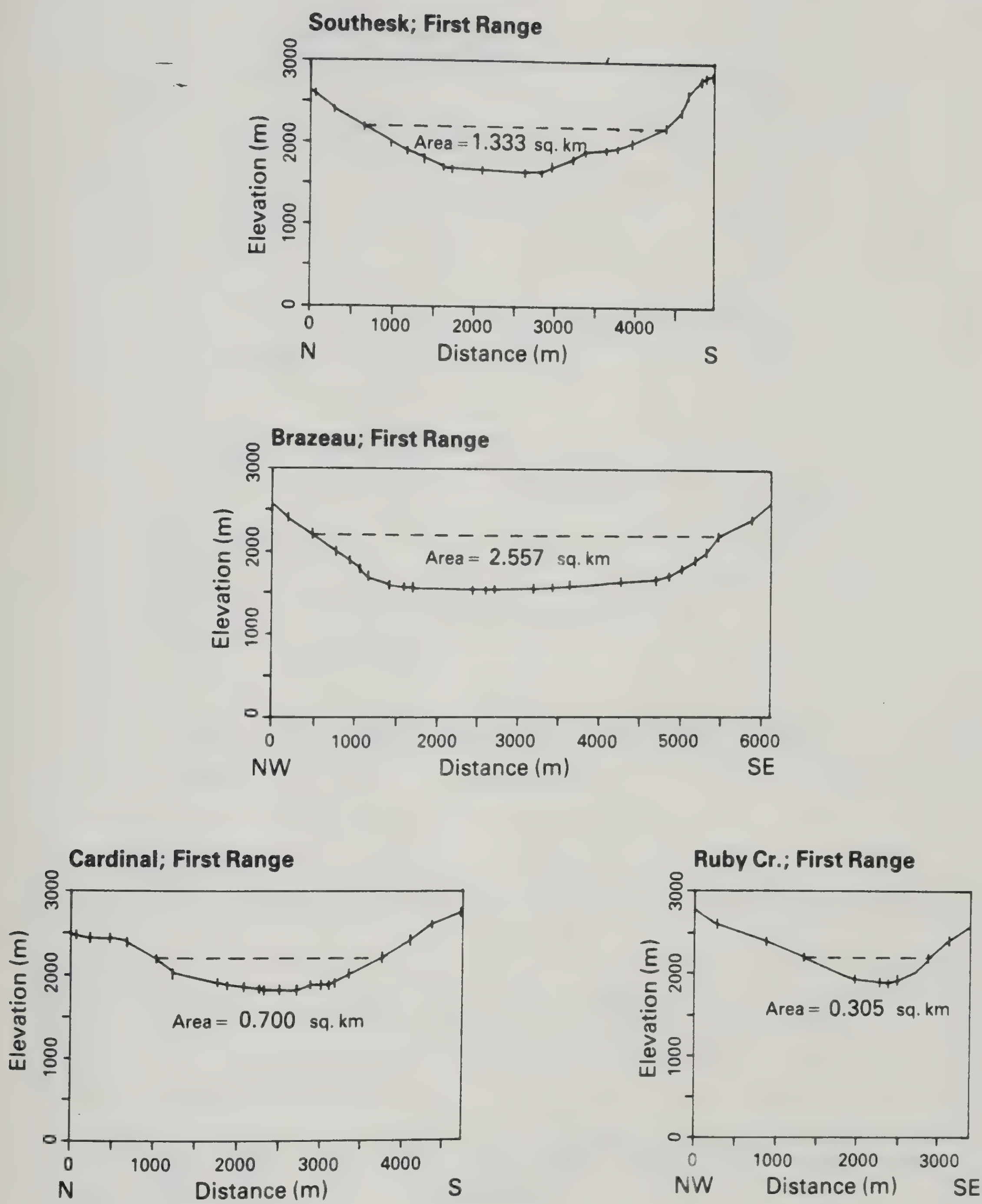

Fig. 5.4 Cross-sections of valleys at First Range

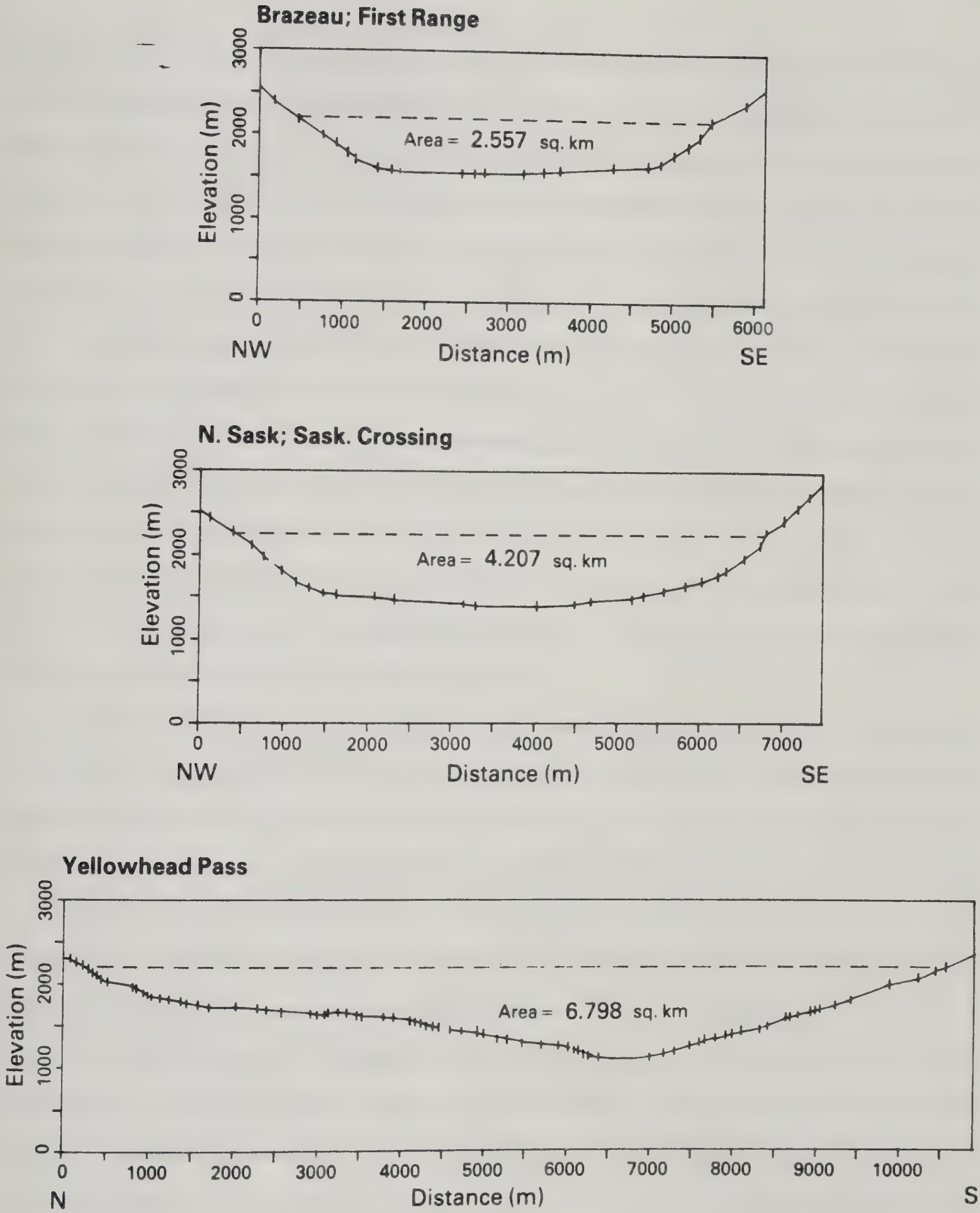

Fig. 5.5 Comparison of cross-section size of the Brazeau valley and the valleys of the North Saskatchewan and Yellowhead

location, and it is more than one-third the size of the glacial trough at the Yellowhead Pass. This suggests that the principal source area for ice of the Brazeau Lobe must have been from within the Front Ranges rather than further to the west as was the case in these other two valley systems (see Figure 5.5). The landforms, discussed in Chapter 3, furnish very little unequivocal evidence of ice margins within the study area. Lateral and end moraines are sparse even when it is considered that a conservative approach was taken to identifying such landforms. For example, the hummocky moraine immediately west of the confluence of Crooked Creek and the Brazeau River valley may well be a poorly-preserved lateral moraine which wraps around the bedrock ridge and extends into the Crooked Creek valley. Also, partly infilled portions of structural valleys along the southern rim of the Brazeau River valley may be remnants of a lateral moraine system. If so, it could be argued that there is evidence of an ice-marginal landform suite at this level and this would support the explanatory model of multiple glaciations proposed by Reimchen and Bayrock (1977), as well as others.

Well-preserved, but small, lateral moraine remnants exist along the west side of the Forestry Trunk Road Valley (Figure 3.1a) and north of the Brazeau River valley on the southeast flank of the Cardinal Hills. The morphology and topographic position of these features (see Figure 5.1) indicate the former presence of glacier ice within the main valley at the eastern side of the Foothills. The only possible sources of the ice which formed these features were the Athabasca Lobe, which occupied the Western Alberta Plains, or the Brazeau Valley Lobe which flowed from the First Range to the west. If the ice source was from the northwest (Athabasca Lobe), one would expect to find extensive glaciolacustrine deposits within the Brazeau River valley to the west beyond the point of maximum possible penetration of the Athabasca ice. Because such sediments are generally absent in that area, this is considered to be an unlikely explanation. The Landforms Map quite clearly illustrates that, although there are extensive glaciofluvial and glaciolacustrine deposits along the eastern slopes of the Foothills, the landforms within the Brazeau River valley are predominantly of glacial and glaciofluvial origin, from the Front Ranges of the Rocky Mountains to the eastern edge of the Foothills. It follows that either the lateral moraines were formed during a glacial event which pre-dates the majority of other glaciagenic landforms within the Brazeau River valley or they were

formed by a glacier that issued from within the Brazeau drainage basin and, for a period of time, abutted against ice of the Athabasca Lobe. The evidence which supports the second alternative is as follows:

(1) Meltwater channel segments, which indicate that meltwater discharges flowed approximately parallel to the Brazeau River valley, were identified immediately south of the main valley. The presence of glacier ice in the main valley, providing topographic constraints, seems necessary to explain why water was ducted across the Foothills ridges in that area.

(2) The orientations of ice-moulded landforms imply that ice from the Brazeau River valley diverged into the Forestry Trunk Road valley and possibly into the Crooked Creek valley at a time when ice levels were somewhat higher than the thresholds of these tributary valleys.

(3) The esker complexes in the Brazeau and Crooked Creek valleys are evidence of stagnant ice wastage within these valleys. In the case of the Thunder Lake esker complex it is improbable, and for the Crooked Creek Valley esker it is nearly impossible, that these landforms were formed from ice sources which entered the valleys from the east.

(4) No evidence of end moraines related to the main valley glacier was found in the Brazeau River valley, within either the Foothills or Mountain regions. In fact, there is no evidence of end moraines beyond approximately two kilometres of the present Brazeau Icefield.

Glaciofluvial and glaciolacustrine landforms indicate areas where ice provided both water and sediments. Where the glaciers altered drainage patterns, these landforms are useful indicators of former ice-marginal positions. Two anomalous suites of lacustrine and glaciofluvial landforms were observed. Their topographic position shows that these landforms were created in ice-marginal environments. Along the eastern margin of the Foothills extensive glaciofluvial and glaciolacustrine deposits were mapped. Thin mantles of glaciolacustrine material were noted but not mapped over a much wider area. The present land surface slopes toward the east-northeast, transversely to the orientation of most of the meltwater channels, the eskers south of the Brazeau River, and the band of glaciofluvial and glaciolacustrine deposits which are distributed along the

eastern slopes of the Foothills. It is therefore highly probable, and generally accepted, that glacier ice occupied the adjacent parts of the Western Alberta Plains and obstructed the drainage emerging from the drainage basins to the west. The probable time of this occurrence will be discussed later.

In the Foothills southeast of Thunder Lake there is a large area mantled with lacustrine deposits. These deposits were traced to elevations of at least 1500 m; considerably higher than the floors of the structural valleys in the area. A glacier occupying the Brazeau River valley must have obstructed drainage to the east to allow the formation of this ice-marginal system. The glaciolacustrine deposits clearly indicate the presence of ice within the Brazeau River valley. If these deposits date from a pre-Late Wisconsin advance, one would expect to see fossil ice-wedge structures in them because they do occur in the plains at lower elevations and in the intermontane region of Wyoming (Mears, 1981) which were probably subjected to a less severe climate during the last glacial advance. This hypothesis is also supported by the presence of discontinuous meltwater channel segments which transect the Foothills ridges from Coast Creek, across Blanchard and Moosehound Creeks, to the Forestry Trunk Road valley. These appear to have moved water parallel to the present Brazeau River but at an elevation of approximately 1450 m. Similarly, a fluvially-cut notch transects the eastern Foothills ridge northeast of Brown Creek. The glaciofluvial deposits associated with this feature are graded toward the northeast, normal to the strike of the Foothills. It is difficult to imagine how water could breach this ridge unless ice obstructed all topographic outlets and the regional gradient was approximately from west to east. The preservation of glaciofluvial deposits to the east of the notch indicates that an event of glacial erosion or deposition has not occurred in this area subsequent to the deposition of the glaciofluvial sediments.

5.2.1 AREAL EXTENT OF ICE WITHIN FOOTHILLS

The areal extent of glaciation, within the Foothills, between the major river valleys, is controversial. Reimchen and Bayrock (1977), and others, feel that the higher surfaces of the Foothills remained unglaciated since at least Early Wisconsin time. Although there are few, if any, local glacial troughs, ice-moulded moraines, hummocky or kettled

moraines, and no cirques, aretes or horn peaks within the Foothills of the study area, there is still evidence that they were covered by ice during the same event that formed the glacial landforms within the major valleys. Essentially, the evidence is as follows:

(1) Erratics and ice abraded bedrock were observed on Red Cap Mountain and along the Big Horn Range at elevations up to about 2075 metres. The adjacent Foothills surfaces are at least 230 m lower.

(2) Limestone and quartzite erratics, though infrequent, were occasionally found in the surficial deposits mantling the Foothills.

(3) Eskers are preserved within the Foothills well away from the main valley. For example, a long esker lies to the north of the Cardinal Hills indicating that at least this area was once ice covered. The age of this feature is not known but is seems unlikely that the sharp-crested, 20 m high ridges would maintain slopes as steep as 32 degrees over several tens of thousands of years. In turn, soil profile development, the formation of weathering rinds on clasts and the formation of calcite horizons, were comparable to those of similar landforms within the main valley.

(4) The main Fairfax Lake sediment core obtained for this study contains bands of coal below the 850 cm level (see Section 4.3.2). These probably reflect a change in the sedimentary environment from a proglacial lake to a nonglacial lake. The pollen evidence tends to support such a claim. If this interpretation is correct active ice must have reached an elevation of at least 1315 m along the eastern edge of the Foothills in order to allow ice-marginal drainage into this lake basin.

(5) The lateral and hummocky moraines which flank the Brazeau River valley indicate that active ice flowed down this valley to the eastern margin of the Foothills. The upper surface of this glacier attained an elevation of approximately 1500 m A.S.L. (see Section 3.4.1). At that stage re-entrant glaciers should have extended well into the adjacent Foothills valleys. The evidence, however, is that the glacier deposited moraines across the junctions of these valleys with no deflection into them except at the eastern side of the Foothills where ice issued into the Crooked Creek and Forestry Trunk Road valleys.

This evidence demonstrates only that these portions of the Foothills have been glaciated. It does not prove that the ice occupied these areas during Late Wisconsin time. It provides significant support for the hypothesis that the ice that occupied these areas was synchronous with the glaciation of the major valleys such as the Brazeau River valley. The explanation offered for the comparatively sparse evidence of glacierization of the Foothills is that in the main, the ice was probably of local origin and relatively inactive. It has been noted that most of the ice which made up the Brazeau Glacier system originated from within the Front Ranges. Local precipitation and temperature regimes may also have resulted in ice accumulation within the adjacent Foothills. The Foothills ice was thin, possibly cold-based, and the ice surface gradients may have been insufficient to generate significant flows. As mentioned earlier, the maximum surface elevation appears to have been in the order of 230 m higher than the ridges adjacent to Red Cap Mountain and the Big Horn Range. This is considerably thinner than the ice depths which are thought to have occupied the major valley systems such as the Cardinal, Southesk and Brazeau river valleys, or the ice which breached Nigel Pass, one of the shallowest parts of the glacierized Brazeau Valley. It is therefore possible that the ice was thin enough to allow basal temperatures to drop below the pressure melting point, at least along the ridge crests. Even if the postulated ice overlying the Foothills was not cold-based, a necessary condition for the movement required to produce glacial erosion is a shear-stress sufficient to produce flow in the glacier. If the glacier was thin and the roughness of the bed relatively large, as the evidence indicates, the local ice within the Foothills would have been confined by the structural ridges. It would also have been constrained from flowing accordant to the structural ridges by the glaciers which occupied the major transverse valleys extending from the First Range. A final factor contributing to the poor evidence of active Foothills glaciation beyond the major transecting valleys could be the recessive nature of the local bedrock. The soft sedimentary rocks which comprise the bedrock units weather rapidly under cool, moist conditions, so that the preservation of small-scale glacier erosional evidence may have been short-lived.

5.3 DIRECTIONS OF ICE FLOW WITHIN THE STUDY AREA

Ice flow indicators include ice-moulded landforms, abraded rock protrusions and eroded troughs. Such geomorphic evidence indicates that two major ice streams, the Brazeau Lobe and the Athabasca Lobe, traversed the study area. The flow directions illustrated by these landforms are discussed in Section 3.4.1 and generalized in Figure 3.3. The clearest examples of ice flow indicators are also depicted on the Landforms Map. Generally, the pattern indicates that ice of the Athabasca Lobe flowed southeast across the area east of the Foothills. East of the study area the Athabasca Lobe was, for a time, in contact with the Laurentide Ice Sheet. The contact zone between the Laurentide and Cordilleran ice lies along a line, approximately parallel to the Foothills, which transects the Brazeau River valley immediately east of the Brazeau Reservoir. Because ice-marginal or interlobate landforms are poorly preserved the precise boundary can only be approximated. However, glaciofluvial and glaciolacustrine sediments clearly signify the general area where the two ice masses locally separated during initial deglaciation.

The western margin of the Athabasca Lobe appears to have been partly controlled by the easternmost Foothills ridges, and transected the Pembina and Brazeau River valleys east of the Foothills. Remnants of lateral moraines, ice-moulded landforms and a small esker, indicate that at some late stage of glaciation, Brazeau Lobe ice was deflected southeastward along the Forestry Trunk Road valley and northeastward into the Crooked Creek valley. South of the Brazeau River valley, and east of the Foothills, the ice-moulded landforms indicate that the Athabasca Lobe abutted against the eastern slopes of the Foothills an unknown distance beyond the study area toward the North Saskatchewan River. The fact that ice of the Athabasca Lobe did not deflect significantly eastward as it crossed the Brazeau River valley, but continued to flow slightly in toward the Foothills instead of away from them, as local topography would dictate, attests to the relative magnitude of the Athabasca Lobe. It also indirectly reflects the lateral constraint imposed on this lobe by the Laurentide ice sheet further east. No evidence of easterly flow directions was observed from landforms of the Athabasca Lobe in the study area. It is assumed, then, that the Athabasca Lobe was laterally constrained by Laurentide ice at least during the time(s) of ice-moulded landform development.

6. DISCUSSION AND CONCLUSIONS

6.1 INTRODUCTION

The glacial history of the Rocky Mountains and Foothills in Alberta continues to be one of the most controversial of Quaternary subjects in this region. This is partly related to the problem of generalizing information from studies carried out by different researchers working in separate drainage basins. Shaw (in Slaymaker and McPherson, 1972) points out some of the difficulties inherent in such correlations. Most of the studies pertaining to the Pleistocene glacial history of western Alberta have placed the major emphasis on stratigraphic evidence and only a minor amount of attention has been given to the landform assemblages produced by the glaciers. For example, Roed (1968, 1975) briefly described the Hinton terraces, the Marlboro delta, the Emerson Lakes esker and the extensive ice-moulded landforms which radiate from the Athabasca Valley. Boydell (1970) made few references to the glacigenic landforms found in his study area. Obviously, the geomorphic evidence was not ignored because both of these authors obtained much of their information from aerial photographs, but the geomorphic evidence received fairly terse treatment in their reports. Unquestionably, stratigraphic evidence is vital to any interpretation of the glacial history of an area, but models of glaciation and deglaciation ideally should explain the distribution of landforms as well as the stratigraphy of the surficial deposits. The present study area is typified by surficial deposits which are commonly thin or poorly exposed while topography is highly variable. Therefore, of necessity, a disproportionately large emphasis is placed on landform characteristics and distribution patterns in this study.

6.2 MODELS OF GLACIATION AND DEGLACIATION

There appear to be four explanatory possibilities relating to the events of the last major glaciation to affect the area.

(1) There was glacier ice within the Brazeau River valley but no ice in the Foothills or on the Western Alberta Plains.

(2) There was no ice in the Foothills portion of the Brazeau River valley when the

Athabasca Lobe extended down the eastern side of the Foothills.

(3) There was active glaciation of both the Brazeau River valley and the Western Alberta Plains during the last glacial event but the two events were non-synchronous.

(4) The Brazeau Lobe and the Athabasca Lobe coalesced during the last glacial event in the area.

If the first explanation is assumed it should be confirmable by ice-marginal landforms and sediments, at least at the point of maximum extension. Conspicuous Holocene moraines associated with contemporary glaciers, and well-preserved end moraines of Late Wisconsin age from southwestern Alberta and the Rocky Mountains south of the international border, indicate that it is realistic to expect these features to have been preserved. As discussed earlier (Section 3.4.1), there is some physical evidence to support this model. However, a weakness lies in the fact that this cannot satisfactorily explain some of the landform assemblages. Most of the glaciofluvial and glaciolacustrine evidence in the upper Foothills valleys would then necessarily relate to an earlier glacial event. None of the relative-age dating techniques employed suggests that this is probable.

If the second explanation is assumed, confirmation in the form of ice-marginal features along the eastern slopes of the Foothills and transecting the major river valleys, such as the Pembina and Brazeau, would be expected. In addition, ponding resulting from obstruction of local drainage from the Foothills would be expected to produce extensive ice-marginal lakes. Sections B9 - B16 (Sec. 4.2.3) clearly illustrate that glaciolacustrine deposits are common east of the Foothills but the Landforms Map shows that moraines and eskers occupy the floor of the Brazeau River valley within the Foothills. The absence of a lacustrine mantle over these landforms suggests that extensive ponding did not post-date the formation of these landforms. glaciolacustrine sediments were observed along the floor of the Brazeau River valley. Lateral moraines were found along the eastern Foothills but, as discussed above (Section 5.2), these most probably relate to ice of the Brazeau Lobe rather than the Athabasca Lobe. In addition, no geomorphic evidence of re-entrant ice extending into the structural valleys of the Foothills was observed. Contemporary glaciers which occupy main valleys but not the adjacent tributary valleys show extensive re-entrant moraines.

The third explanation is attractive in that the respective drainage basin characteristics, and glacier response times of the Brazeau and Athabasca systems were probably not very similar. Therefore it is plausible that the maximum extensions of both lobes were not time-synchronous. A weakness of this model, similar to those of the two previous explanations, is that if the two ice lobes were non-synchronous the physical evidence would be the same as if they were two separate events. The landform evidence provides very little clear evidence to support such a model.

The fourth alternative provides the most complete explanation of the landform assemblages mapped in the study area. Confirmation of this model could be accomplished in part by regional comparisons. The landform assemblages in the adjacent areas might be checked to determine whether or not this model could serve to explain the landforms in those areas. The applicability of this model to the Athabasca and the North Saskatchewan valleys will be discussed later. For the moment it is sufficient to say that the model will serve reasonably well for the Athabasca Valley and has some applicability for the North Saskatchewan Valley.

Evidence which could falsify the model might include demonstrable differences in the relative age of deposits in different parts of the study area. Reimchen and Bayrock (1977) attempted to use the depth of carbonate leaching as an age-dating technique but the results are considered to be questionable. Nested suites of landforms which are clearly related to multiple glacial events have been frequently employed to show sequential glacial events. Such suites have not been identified within the study area. Stratigraphic evidence might falsify the model, for example, if overlapping sediments of two different provenances were found. This would imply that sequential events occurred at some location(s). Similarly, non-glacial sedimentary units separating tills or other glacigenic deposits could at least suggest a more complex glacial history than is implied by the geomorphic evidence. Such stratigraphic evidence appears to be absent in deposits of the study area. Because the fourth model has not been falsified, and seems sufficiently comprehensive to explain the observed evidence, it is recommended that it be provisionally accepted.

6.3 CHRONOLOGIC EVIDENCE

A number of techniques have been used by previous researchers to determine the age(s) of the deposits along the eastern slopes of the Rocky Mountains. It is generally accepted that radiometric techniques provide the greatest degree of accuracy but unfortunately extensive searches by researchers in the Rocky Mountain Foothills have disclosed very little organic material within or beneath the glacigenic sediments. No datable material was found within or beneath any of the glacial units in the study area. Lake and bog cores were considered to at least provide minimum ages of deglaciation. The bog cores produced basal dates of mid-Holocene age however (see Table 4.1). The young basal dates may be explained in a number of ways. A steel probe was used to determine that the cores represented the entire bog stratigraphy but it is possible that the corer did not reach the lowest organic sediments. The corer and the probe may have been obstructed by a beds of mineral material such as loess or aeolian sand. Alternatively, it is possible that the dates are valid and that the organic accumulations in this area were oxidized or eroded away during the drier Altithermal episode and recommenced later during the return to cooler and moister conditions. The general agreement of these dates with others from west-central Alberta lends support to the latter explanation. A third explanation is that the bedrock underlying these areas is sufficiently permeable to allow humic acids to migrate down through the sediments, contaminating the lower deposits. This is not very likely as the substratum must be relatively impermeable in order to develop an accumulation of organic sediment.

Two lakes within the study area were found suitable for coring. They were sufficiently deep that they probably persisted throughout the Altithermal phase and they have no significant streams draining into them. Thus their sedimentation rates would have been slow. Radiocarbon dates were obtained from both Fairfax Lake and the Muskiki Lake cores. The Fairfax core date of 11,225 +/- 120 years B.P. (S-1705), along with stratigraphic and palynologic evidence, suggest that the lake was initially an ice-marginal system which evolved into a lake sustained only by a local catchment area. The dated section of the core is considered to be a useful minimum date for local deglaciation. Ice probably remained within parts of the study area later than this time but available dates from sites west of the study area (Westgate and Dreimanis , 1967; Luckman and Osborn,

1979) of about 10,000 years B.P. cast doubt on the validity of the 8300 year old age assigned to basal sediments of Muskiki Lake. It is concluded that wastage of Late Wisconsin ice exposed much of the study area sometime prior to 11,000 years ago. The relative-age dating evidence, although only superficially tested, tends to support this conclusion. No nested ice-marginal landforms were found between the presumably Holocene moraines in the headwaters of the Brazeau River valley and the Brazeau Reservoir. No significant differences in soil profile development or weathering depths were observed within the study area. No geomorphic boundaries were observed beyond which the landform assemblages generally appear to be more subdued and drainage networks better developed. Any of these lines of evidence would tend to falsify the model, but despite a serious investigation to produce such evidence, none was found. It is therefore suggested that the following conclusions be made.

(1) The largest lobe of ice to effect the study area flowed southeast along the front of the Foothills from the Athabasca Valley.

(2) It is probable that the Brazeau Lobe and the Athabasca Lobe coalesced for a time at, or near, the eastern edge of the Foothills.

(3) Most of the ice which issued from the Brazeau Valley originated within the Front Ranges. The erosional evidence indicates that the amount of ice that entered the Brazeau Valley from other valley systems further west was relatively minor.

(4) Deglaciation was largely by area-wasting rather than by frontal retreat.

(5) Late Wisconsin ice exposed the eastern Foothills, within the study area, at about 11,225 years B.P. or somewhat earlier.

(6) No evidence was found to support the case for subsequent less extensive advances of valley glaciers in this area except for the minor Holocene advances near the valley heads.

The model proposed by this study is thus relatively simple. During the late Wisconsin glaciation mountain glaciers formed in the Brazeau Valley system with only a minor direct contribution of ice from the Cordilleran complex to the west of the Continental Divide. While only a small amount, if any, of the Cordilleran ice from west of the divide was added to the Brazeau Lobe, a larger amount issued east of the Foothills via the Athabasca Valley. Laurentide ice, lying further to the east near the present site of the Brazeau

Reservoir, deflected the Athabasca Lobe southeastward sub-parallel to the topographic trend of the Foothills.

During at least the later stages of this advance the Athabasca Lobe merged with the ice of the Brazeau Lobe which issued from the Brazeau River valley. The two ice masses coalesced at the eastern side of the Foothills. The evidence for the adjustments in ice flow produced by the confluence of the two ice bodies indicates that only minimal deflection occurred in the Athabasca Lobe. This is probably partly due to the fact that the eastern ridge of the foothills (Cardinal Hills) obstructed the flow of the Brazeau glacier which forced some distributary ice to flow northward into the Crooked Creek Valley and southeastward into the Forestry Trunk Road Valley.

From this maximum phase both ice ice masses seem to have rapidly downwasted. There is some evidence that during this ablation phase the less active ice within the Foothills released meltwater to the east across the structural ridges of the Foothills. As the ice surface lowered the drainage was initially controlled by ice topography and the meltwater networks radiated away from the main valley. Evidence for this is particularly striking toward the southeast portion of the study area. On the Western Alberta Plains some of the channels indicate a flow to the east and southeast away from the Foothills. Many, however, are oriented sub-parallel to the ice-moulded topography. These channels do not appear to have been deflected by the ice-moulded ridges so it is reasonable to suppose that they were superimposed by the wasting Athabasca Lobe. Later, it appears that for a time the Athabasca Lobe obstructed meltwater drainage to the east of the Foothills. Ponding was extensive on either side of the present Brazeau Valley. It seems likely that Fairfax Lake came into existence soon after this, probably a little earlier than 11,000 years B. P. Figures 6.1 to 6.5 illustrate, in a general way, the interpreted phases of deglaciation outlined above.

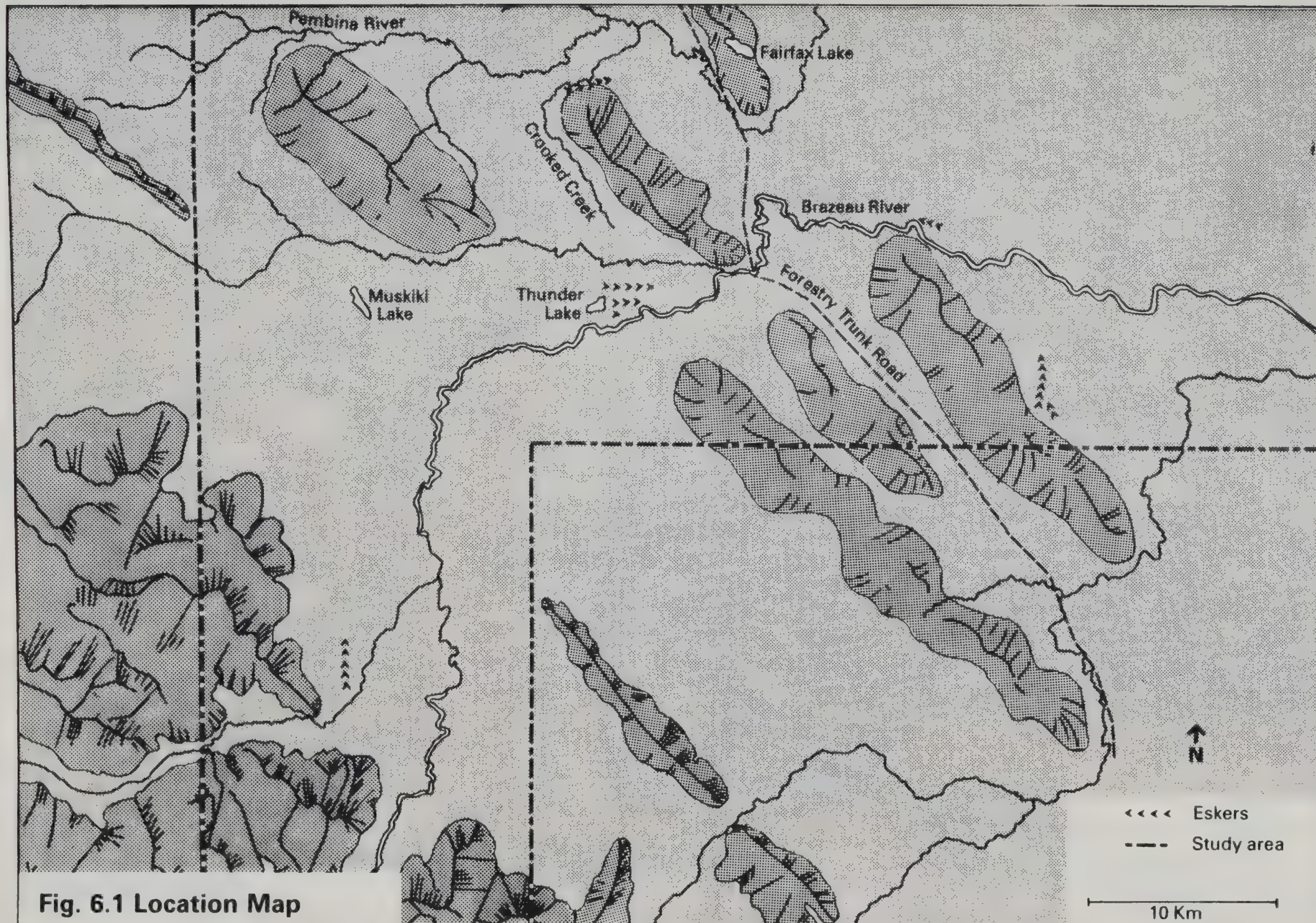

Fig. 6.1 Location Map

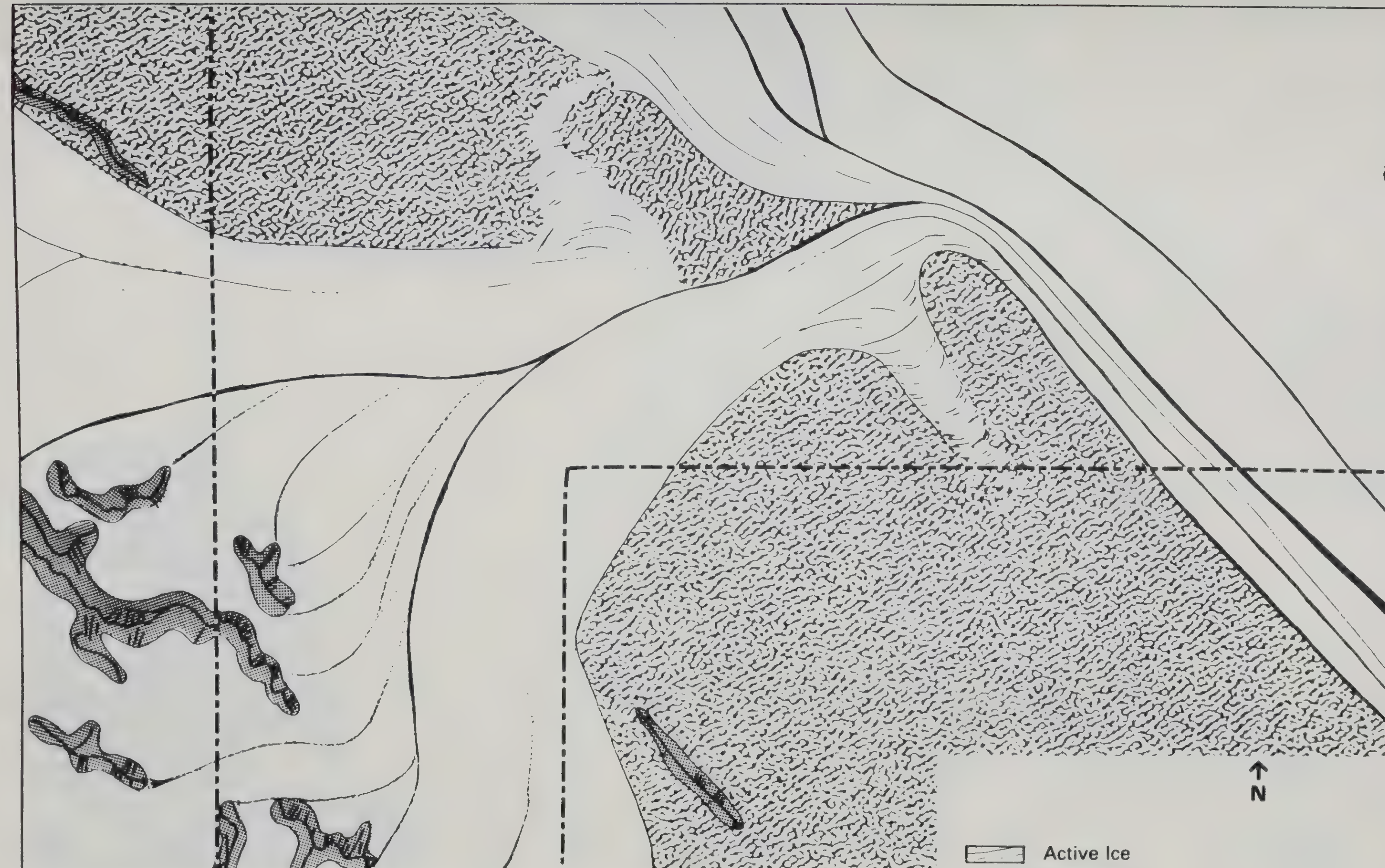

Fig. 6.2 Maximum Glaciation: Brazeau Glacier confluent with Athabasca Glacier

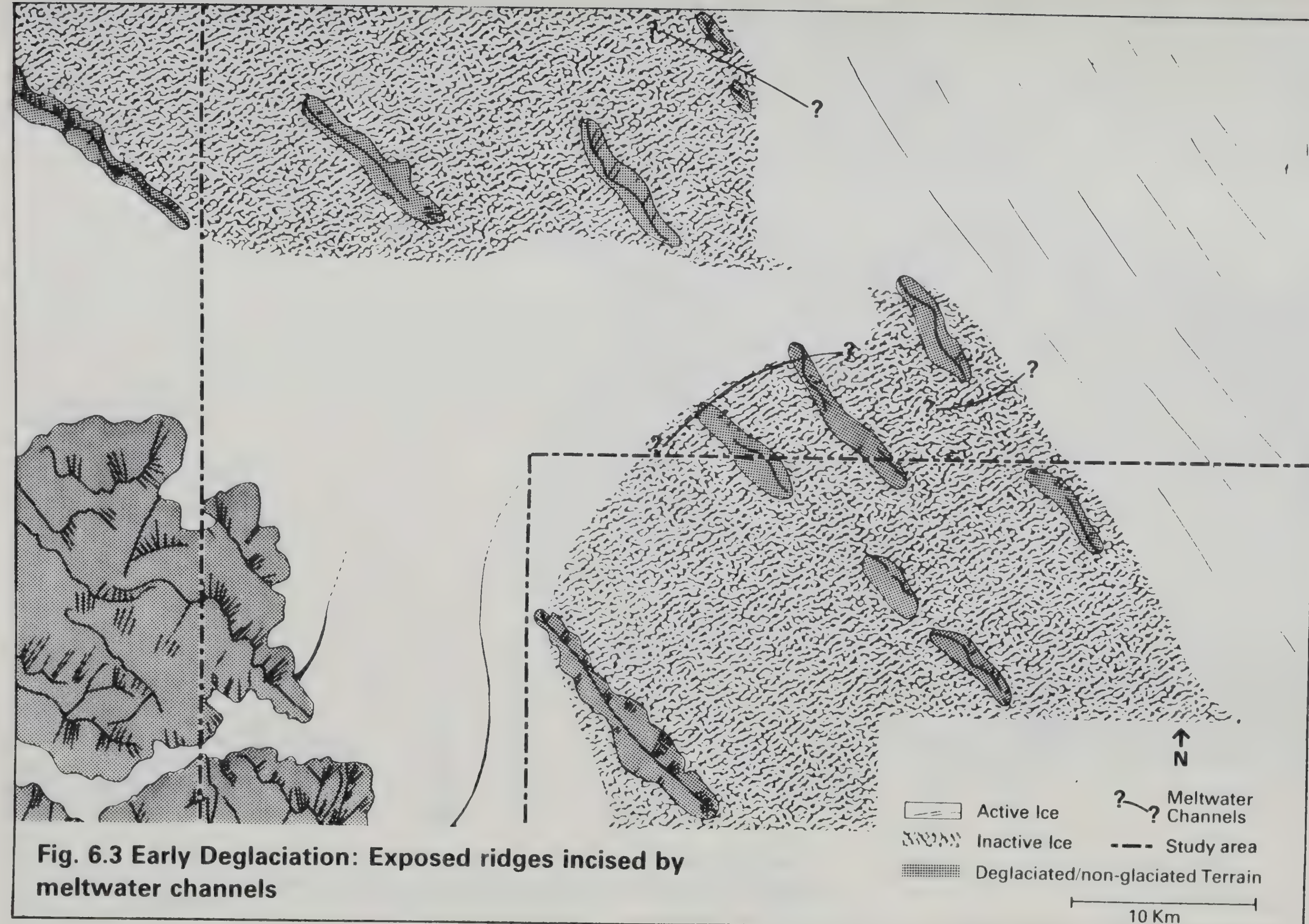

Fig. 6.3 Early Deglaciation: Exposed ridges incised by meltwater channels

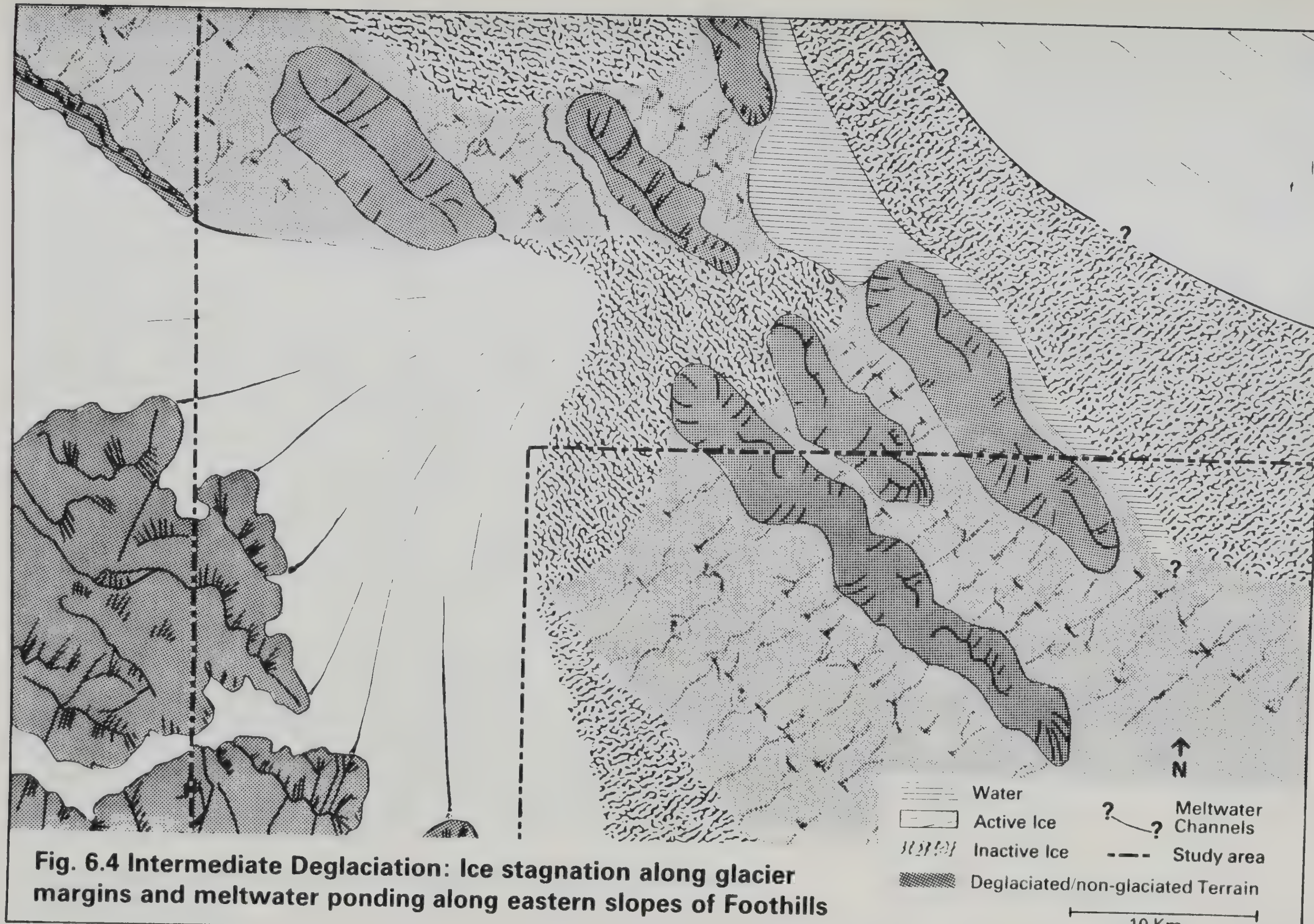

Fig. 6.4 Intermediate Deglaciation: Ice stagnation along glacier margins and meltwater ponding along eastern slopes of Foothills

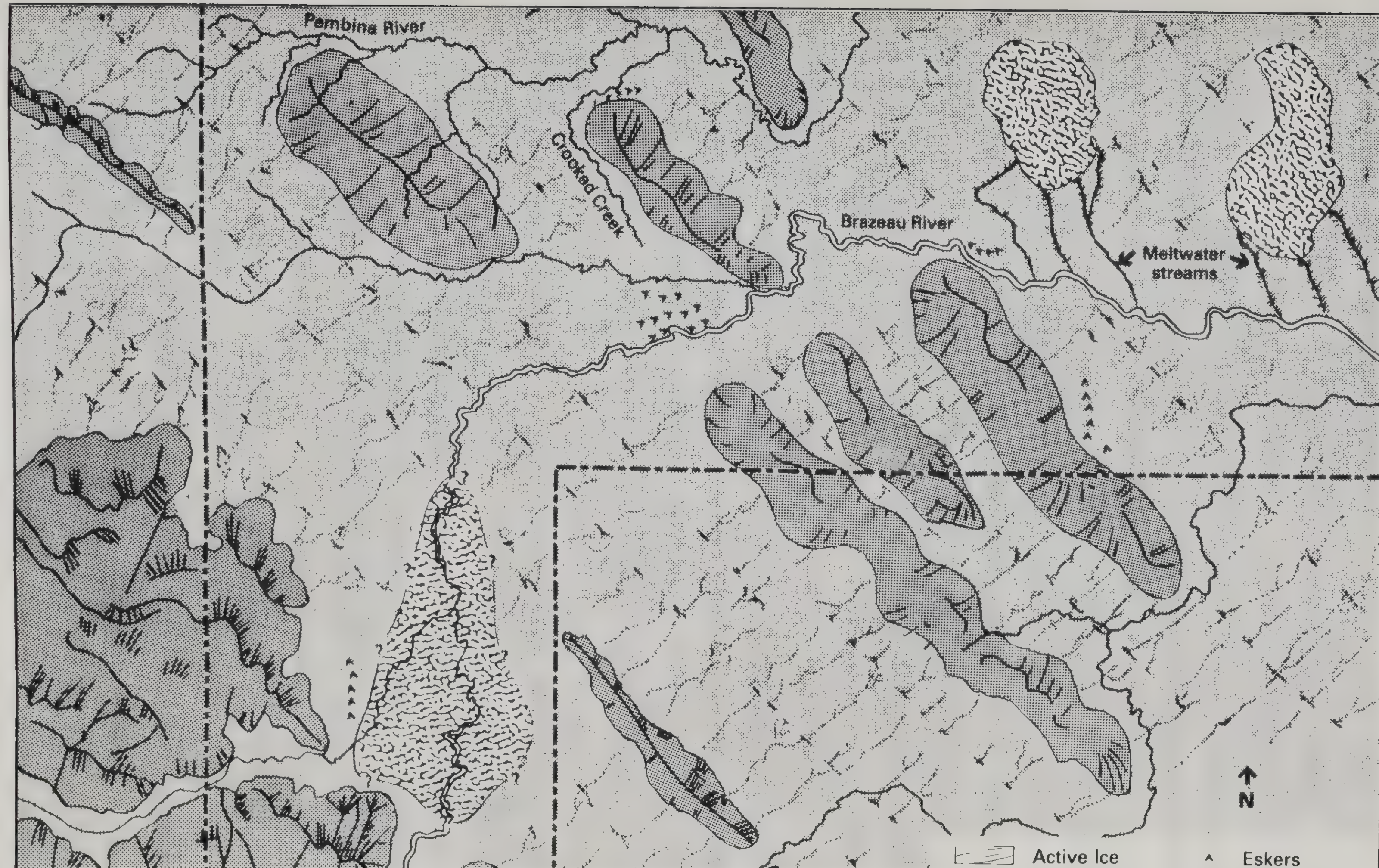

Fig. 6.5 Late Deglaciation: Isolated blocks of ice ablate from lower elevations

6.4 REGIONAL RELATIONSHIPS: A REVIEW

Perhaps in response to pressure from anthropologists wishing to explain the migration of early man into North America, or in order to extrapolate glacial histories from south of the Canada–United States border, there has been a tendency to group the accumulated evidence of glacial episodes into three Cordilleran and one or more Laurentide advances. Reeves (1973) and Rutter (l978), 1980, in Rutter and Schweger, 1980) have synthesized much of the available literature on the topic and find general support for this model. It is also presented in Reimchen and Bayrock (1977) for this study area and in a paper by Stalker (1977) which deals principally with Laurentide ice margins. Accordingly, the model incorporates the idea that the Foothills of Alberta, except for the major river valleys which transect them, have been ice-free for more than 50,000 years.

This model is based primarily on the superimposition of informal litho-stratigraphic units. Few radiometric dates are available for these units. Confirmation of this model has developed mainly by correlating the observations on litho-stratigraphic units between valley systems. This is a useful exercise because it assists in the development of a regional glacial model but problems persist. For example, Sugden and John (1976, p. 234) made a pertinent observation in their discussion of stratigraphic evidence;

> "A final fact to be borne in mind in connection with the above paragraphs (*about sedimentology and structures in till*) is that till is seldom deposited in isolation. In subglacial and englacial situations, fluvioglacial materials may accumulate adjacent to dirty ice which is evolving into melt-out or lodgement till. It is not unknown for lodgement till and lacustrine deposits to occur side by side. On downwasting glacier surfaces especially, fluvioglacial and lacustrine materials may be common, and they will have complex spatial and stratigraphic relations with melt-out tills and flowtills. Sequences of deposits in which there may be several tills interbedded with water-laid deposits are common in ice-wastage environments of today; and indeed (...*it can be demonstrated* ...) how very complex bipartite or tripartite till sequences with interbedded sands and gravels may result from a single phase of glacier wastage....Boulton (1967) warned against the facile assumption .... that every till

represents a glacier advance and every layer of fluvioglacial material a glacier retreat."

It has become generally accepted that ice stagnation was common along the Eastern Slopes of the Rocky Mountains during the deglaciation phase of the Late Wisconsin event (McPherson, 1970; Rutter, 1972; Luckman and Osborn, 1979). This increases the demands on models based largely on stratigraphic evidence as the surficial deposits from stagnant ice are expected to be more complex than if frontal retreat was the dominant pattern of deglaciation. Firmer confirmation of this model may be established when additional radiometric dates become available from critical sedimentary units. Even a series of minimum dates, such as that presented in this study, should give some indication of surfaces which pre-date the Late Wisconsin advance. Basal dates and/or dates from above and below major beds would, of course, be the most desirable and powerful means of confirming the concept.

Another form of confirmation may lie in the geomorphic evidence. This is generally only useful for the most recent event as subsequent glacial events usually obliterate or alter the landforms from prior episodes. The need to employ geomorphic confirmation of the stratigraphic model can be illustrated by the evidence from the Athabasca Valley. At the Continental Divide (Yellowhead Pass) the glacial event that produced the kettled topography and lacustrine sediments described by Smith (1979) must have been associated with the Obed Advance, or perhaps the Drystone Creek Advance of Roed (1968, 1975). If these deposits were formed by the earlier and more extensive Marlboro Advance (Roed, 1975) it is difficult to imagine why the ice of the Obed Advance would flow east beyond the mountain front but fail to occupy the Yellowhead Pass area.

The radiometric dates presented by Harrison (1976) Clague (1981) and Ferguson and Osborne (1981) indicate that ponding occurred along the western side of the Continental Divide in southeastern British Columbia between about 12,000 and 14,000 years ago. This supports the contention that the glacial advance responsible for the ponding took place during the Late Wisconsin time. Although it weakens the argument considerably to extrapolate radiometric dates over such distances, this lends support to the hypothesis that Late Wisconsin ice did cross the Continental Divide.

Evidence which might tend to falsify the model of multiple glaciations would include radiometric dates which are inconsistent with the age assigned to the surface. Similarly, if future stratigraphic work indicates that the previous interpretations were incomplete or erroneous the model will be falsified or will have to be modified to take the 'new' evidence into account. If the geomorphic evidence is inconsistent with interpretations based on stratigraphic evidence, the stratigraphic models are not falsified, but rather they are not confirmed and a need for further research is identified. The importance of this distinction is that if secondary evidence fails to confirm the hypothesis the hypothesis is weakened (i.e. it is less likely to be true) but if the hypothesis is falsified, the model is rejected and the evidence must be reviewed and a more appropriate model must be proposed.

Some of the models of glaciation that have been proposed for the study area are as follows: Reimchen and Bayrock (1977, p.18) in their study of the entire Alberta section of the Rocky Mountain Foothills, proposed that:

> "The Classical Wisconsin glaciation in the Rocky Mountains during the maximum extent of glaciers has the pattern of valley glaciation of the dendritic form. In contrast to this, the Intermediate till was deposited from glaciers approaching the reticulated form, and the Old tills were deposited most likely from a mountain ice sheet through which numerous nunataks were protruding."

Later, in the same report (Reimchen and Bayrock, 1977, p.43), they elaborate on this:

> "During the latest glacial maxima, over 90% of the area between major mountain valleys in the Rocky Mountains and Foothills was ice free, as can be attested to be (*sic*) the lack of glacial deposits. Several minor resurgences of these valley glaciers are evident as end moraines can be found at successively higher elevations in most of the major mountain valleys. In general, up to four minor readvances of these small alpine glaciers can be shown with the latest glacial advance being confined to cirque basins."

Rutter (1980, p. 3) states:

"In summary then, (*summarizing the available literature*) the picture that emerges is that there was an ice free corridor at least from the Jasper–Edmonton area southward to the U.S. border during Late Wisconsin time. It extended westward from the limits of the Lethbridge advance to at least the mouths of major valleys of the eastern Cordilleran mountain front in the northern part of the area under discussion, and within the major valleys in the southern part... ."

Stalker (1977) outlined a similar model of glaciation based on Laurentide evidence. Figure 1 in Stalker (1977) indicates that a tongue of Cordilleran ice flowed southeast along the eastern margin of the Rocky Mountain Foothills from the Athabasca Valley. The Athabasca Lobe of ice is required, in Stalker's (1977) interpretation, to explain the decrease in elevation northward of a Laurentide end moraine along its western margin. He draws a suggested western boundary of the Athabasca Lobe of Cordilleran ice but does not explain the geomorphic evidence upon which this proposed limit is based, except to say that (Stalker, 1977, p.2617):

"if the Laurentide ice of the Innisfail Lobe had been free to spread westward to an altitude commensurate with its surface elevation, its western limit would have been close to this (*the western limit of Cordilleran ice*) line".

Stalker's (1977) model should not be too harshly criticized for its treatment of Cordilleran ice, because the primary focus of his paper was to define the limits of Late Wisconsin Laurentide ice. However, other researchers (Canby, 1979; Rutter and Schweger, 1980) have used this part of Stalker's (1977) model to support the model which includes a Late Wisconsin ice-free Corridor in part of western Alberta. It is therefore important to point out that the Late Wisconsin Cordilleran Ice mass must have been rather peculiar if it produced a lobe of ice which issued from the Athabasca Valley, deflected 90 degrees because of interference from Laurentide ice, and then extended another 200 kilometres along the eastern edge of the Rocky Mountain Foothills while the ice in other major valleys, such as the North Saskatchewan, only extended down-valley as far as the mountain front.

Perhaps the argument should be re-directed toward the position held by some (eg. Roed, 1968; Reimchen and Bayrock, 1977; Rutter, 1980; and possibly Boydell, 1978)

that the advance which ended 10,000 to 13,000 years ago only reached the eastern edge of the mountains or issued a short distance into the Foothills from the Cordillera. Evidence to support this model is predicated largely on the following:

(1) Rutter (1977) identified a Late-Portage Mountain Advance, evidenced by a deltaic end moraine, which he has radiometrically dated at 11,600 +/- 1000 years B.P. (I-2244A). This moraine formed at the eastern edge of the mountain front in the Peace River valley.

(2) Roed (1975), after personal communication with Rutter, equated the Obed Advance, which extended 30 to 40 km to the east of the mountain front along the Athabasca Valley, to Rutter's Late Portage Mountain Advance (published later in Rutter, 1977).

(3) Jackson (1977) provided a date for deposits of his Erratics Train Glaciation, near Midnapore, of 49,400 +/- 1000 years B.P.(GSC-2409). Peat from a bog near Turner Valley was also dated at 17,500 to 18,400 years B.P. (GSC-2668). His interpretation was that Late Wisconsin Ice did not invade these areas. Jackson's (1977) Erratics Train Glaciation has been tentatively equated with Rutter's (1972) Bow Valley Advance which extended beyond the Front Range of the Mountains. The two subsequent advances (Canmore and Eisenhower Junction) did not extend to the eastern edge of the mountains. These are thought to represent the Middle and Late Stades of the Pinedale Glaciation (Jackson, 1977, p.40).

(4) In suggesting a glacial chronology for the southwest corner of the province Stalker and Harrison (1977) stratigraphically relate Cordilleran Advances to the Laurentide chronology and conclude that the Late Wisconsin advance (Waterton IV) terminated west of the mountain front.

(5) Reimchen and Bayrock (1977) found differences in the depth of leaching of calcium carbonate in tills and suggested that these variations indicate differences in relative age. They used this, together with morphological evidence, to conclude that only the earliest of the last three (Wisconsin(?)) advances extended beyond the eastern margin of the Foothills and the last (Late-Wisconsin(?)) advance only occupied the bottoms of the major mountain valleys. Unfortunately the utility of their study is severely hampered by internal inconsistencies discussed earlier.

Perhaps an alternative model is warranted. The Reimchen and Bayrock (1977) study is most directly related to the current study but, as discussed previously, their technique for the relative dating of the deposits is considered to be questionable. Part of the present study failed to confirm their findings of differing depths of leaching and the geomorphic evidence certainly conflicts with their interpretation. Roed's (1968, 1975) work is also particularly relevant to this study because the Athabasca Valley was the source of a lobe of ice which covered a large part of the study area. The related glacial event was Roed's (1968, 1975) Marlboro/Raven Creek advance. No dates are available for this event except through correlations with Rutter's study (published in 1977) of the Upper Peace River valley. Roed (1975) identified two subsequent Cordilleran glacial events, the Obed and Drystone Creek advances. He correlated these with Richmond's (1965) Middle and Late Pinedale Advances. Perhaps some of Roed's (1968, 1975) stratigraphic evidence should not be conclusively accepted as representative of specific chrono-stratigraphic events. For example, he stated that (Roed, 1975, p.1507):

> "... the presence of the erratics indicates that the Obed glacier may simply represent a readvance of the Marlboro glacier."

The landforms he used to define the limits of the Obed advance are not unequivocal on interpretation of related aerial photographs. The "lobate end moraine comprising part of the terminal moraine ..." (Roed, 1975, p.1507) located in TP 54, R 23 may be alternatively interpreted as a complex area of stagnant ice landforms which have been dissected by eastward flowing meltwater streams. The clearest end moraine remnants are lobate units which straddle Oldman Creek but there is little firm geomorphic evidence to suggest that they are other than minor recessional features developed during the general recession of the Marlboro glacier instead of moraines formed by a subsequent (Obed) event.

Evidence of truncation of the ice-moulded Marlboro landforms by Obed ice-moulded landforms was not observed by the author on the airphoto mosaics or on airphotos. However, Roed (1975) states that the evidence is very clear. Because this might provide critical evidence for the Obed event being a second major advance two aerial reconnaissances flights were made over the area but no truncated drumlins or flutings were observed. If the Marlboro and Obed events were two distinct advances there ought to be more than just stratigraphic evidence and truncated active-ice features

to reflect them. For example, ice-marginal or proglacial channels should be present on the Marlboro surface, radiating away from the margins of the Obed end position(s). The most obvious examples of such channels would be where the ice margins were confined by topography. An airphoto study of the streams which issue from the eastern slopes of the mountains, and transect the Marlboro and Obed surfaces, did not show any evidence of changes in valley morphology across this presumed boundary. If ice lobes remained for a period of time in the maximum position as defined, by Roed (1975), the glacier would have obstructed or diverted streams where the ice margins were oriented obliquely or transversely to the local relief. The Obed advance, in particular, should have obstructed or diverted drainage in these streams because the ice limits were then sufficiently low in the valleys that the streams flowing east from the mountain front would have flowed some distance beyond the steep slopes of the mountains before encountering the Athabasca Valley glacier. Such negative evidence does not constitute proof that the Obed and Marlboro advances were in fact a single event but it does reinforce Roed's (1975) cautious statement that they may have been part of a single event.

Evidence for the Drystone Creek advance (Roed, 1975) is particularly tenuous. Surface investigations, extensive aerial reconnaissance and airphoto investigations by the author failed to identify anything that might be interpreted as an independent late Wisconsin advance down the Drystone Creek valley. No end moraine, or anything resembling an end moraine, could be found. A ridge transects the mouth of the Drystone Creek valley but this ridge extends down the Athabasca Valley from the lee-side of a bedrock prominence up-valley from Drystone Creek and it does not deflect as it crosses the mouth of the Drystone Creek Valley. Further down-valley it decreases in size until it disappears to the east of Drystone Creek.

The theoretical basis for a Drystone Creek advance, as described by Roed (1975), is rather questionable. He proposed a glacial advance which was restricted to the very area that other researchers are arguing was essentially ice-free since at least Early Wisconsin time. Roed (1975, p. 1508) stated that;

> "Drystone Creek till occurs in small valleys typically heading in cirques, along the Front Ranges and Foothills of the Rocky Mountains.... *and that....* "Local Alpine glaciers in the Front Ranges and Foothills of the mountains deposited the Drystone Creek till".

It is difficult to imagine why an advance should take place in only the Front Ranges and the Foothills when there is evidence that all other advances originated from further west and all contemporary glaciers are well to the west of this area. Moreover, the Drystone Creek advance, if it existed, must have occurred earlier than 10,000 years B.P. because it is unlikely that ice would have flowed to the lower reaches of the Front Range while the higher areas near the Continental Divide remained essentially ice-free. Luckman and Osborn (1979) published limiting radiometric dates for the deglaciation of two areas in the Main Ranges of the Rocky Mountains; the Tonquin Pass, approximately 95 km west of the present study area, and Castleguard Meadows, approximately 65 km southwest of this study area. A basal date for peat extracted from a bog within a kilometre of a contemporary glacier, is 9660 +/- 280 years B.P. (BGS-465) in the Tonquin Pass and a date of 9600 +/- 300 years B.P. (BGS-490) was obtained for a similar site in Castleguard Meadows.

Either the Drystone Creek event was a local perturbation of the Late Wisconsin advance or the designation is based on a mis-interpretation of the geomorphic and stratigraphic evidence. Few, if any, cirques may be identified within the Foothills, and those in the Front Ranges appear to have end moraines only within a kilometre or two of their headwalls. The elevations and relative positions of these moraines are somewhat similar to those in the Main Ranges described by Luckman and Osborn (1979). This suggests that these moraines are probably of Holocene or late Wisconsin age.

Boydell's (1978) study of the Rocky Mountain House area outlined a glacial chronology consisting of one Laurentide and three Cordilleran advances. The evidence upon which his model was based is comparable to that described in the present study. For example, he (Boydell, 1978, p.28) found evidence that deposits of the Jackfish Creek advance (Cordilleran) and the Athabasca/Sylvan Lake advance (Laurentide) grade into one another with no evidence of superimposition of deposits. His report included two radiometric dates (9670 +/- 140, I-5677, and 10,250 +/- 165, I-5675, years B.P.)

which were interpreted as being minimum dates for Laurentide deglaciation of the area. On this basis he concluded that both the Jackfish Creek and the Athabasca/Sylvan Lake events were of Late Wisconsin age. He also documented (Boydell, 1978, p.23) extensive glaciofluvial and glaciolacustrine deposits along the presumed former contact zone between the Cordilleran and Laurentide ice masses. This agrees with a conclusion from the present study that Cordilleran and Laurentide ice masses were in contact and moving south-eastward during the Late Wisconsin Advance. This is certainly at variance with the previously noted viewpoint that the late Wisconsin advance of Cordilleran ice was restricted to the major mountain valleys. In fact if Boydell's (1978) interpretation is accepted, the area of ice coalescence during the late Wisconsin time extended at least from the Athabasca Valley to south of Rocky Mountain House. A recent study by Mathews (1980) provides additional support for such an interpretation of ice coalescence. He suggested that Mountain-Cordilleran and Laurentide ice coalescence extended well to the north of the Athabasca Valley. Mathews' (1980) study area is centered around Fort St John, British Columbia, and includes the area mapped by Rutter (1977). Radiometric dates pertinent to the events described by Mathews (1980) indicate that initiation of the last major ice advance may be approximated by dates of 27,400 +/- 850 years B.P., (I-4878); 27,400 +/- 580 years B.P., (GSC-2034) and 24,940 +/- 380 years B.P., (GSC-537) years B.P. Sometime later the Cordilleran and Laurentide ice masses merged. Related ice-moulded landforms indicate that flow directions of both ice masses were deflected along the contact zone. Although Mathews (1980) admitted difficulty in resolving one date from a mammoth tusk which appeared too old to fit his model he concluded his study with a discussion of the possibility of an ice-free corridor in the area (Mathews, 1980, p. 21).

> "It appears, however, that the corridor may have been closed for a time during the climax of the last glaciation by local coalescence of the two ice sheets between about 52 degrees 15'N in Southern Alberta (Harris and Boydell, 1972 p. 49) and 60 degrees 10'N in the southeastern corner of the Yukon Territory."

Interpretations of the time of deposition of the Erratics Train, which transects the Brazeau River valley, are also relevant to this study. Most authors seem prepared to

accept that the entire train of erratics, from the Athabasca Valley to northern Montana, was deposited by a single glacial event. If the Laurentide and Cordilleran advances behaved in similar fashions during each glaciation one might argue that each major glaciation that resulted in coalescence of ice might have transported and deposited some erratics and the Erratics Train may be the product of several glacial events. In any case, the principal question to be addressed pertains to the time that other authors have assigned to the deposition of these erratics. Stalker (1956, p. 17) argued that:

> "The erratics, with their great susceptibility to frost action, could not have withstood much of the severe climate that would be associated with subsequent glaciations."

Presumably that argument also pertains to periglacial environments such as would be experienced if ice was covering most of Alberta except for a corridor along the Foothills. Roed *et al.*, (1967) stated that they did not have sufficient information to ascribe an age to the Erratics Train but agreed with Stalker (1956) that these erratics were probably transported and deposited during Late Wisconsin time. Later, on the basis of his work in southwestern Alberta and radiocarbon dates from the Peace River area (Rutter, 1977), Stalker (1977, p. 904) altered his opinion and placed the coalescence in pre-Classical Wisconsin time. In addition, Jackson (1977) dated wood from his Erratics Train till at 49,400 +/- 1000 years B.P. (GSC-2409) and felt that this till was stratigraphically linked to the Foothills Erratics Train.

One unpublished study provides substantial support for the main conclusion of this thesis that the late Wisconsin advance was the last advance of Cordilleran ice which coalesced with Laurentide ice. That study (of D.G. Smith, University of Calgary) was presented at a meeting of the Western Division of the Canadian Association of Geographers in March, 1979. Smith noted an abrupt variation of landforms within the major passes which cross the Continental Divide. He (Smith, pers. comm., 1979) found evidence of ice retreat to the west of the passes indicating that the Cordilleran ice divide was well within British Columbia during the glacial maximum. This evidence takes the form of near-horizontal kame terraces, glacial lake plains, and deltaic deposits which display evidence of ice-contact surfaces on the west slopes of the divides. East of the divides the valleys are characterized by landform assemblages which show no evidence of

meltwater obstruction (i.e. ice marginal meltwater channels, sloping kame terraces and an absence of lacustrine and deltaic sediments). Smith (CAG presentation, 1979) suggested that the relative relief of the passes probably reflects the amount of ice erosion that they have sustained. If this hypothesis is valid the relative relief of the passes along the Canadian section of the Continental Divide indicates that the maximum amount of erosion was sustained by the Yellowhead Pass and the intensity of erosion decreased both to the northwest and southeast of this area. The Brazeau Valley, although less intensely eroded than the Yellowhead Pass area, is nevertheless sufficiently close to the area of maximum throughflow of ice to have contained a substantial glacier, as the geomorphic evidence suggests. Some authors (Rutter, 1965; McPherson, 1970) have stated that the lack of erratics from west of the Continental Divide indicates that the Continental Divide acted as an ice divide during the last major advance. If the Late Wisconsin advance was a mountain valley glacier complex which originated essentially within the present fluvial basins the ice from west of the divide only crossed during some earlier more extensive advance. How is it then, that the lacustrine and glaciofluvial sediments have not been re-worked to the west of the divide? How could ice have formed somewhere in the Rocky Mountains to a level which caused the ice flow to extend at least to the eastern mountain front and yet no ice flowed westward from the divide?

Unfortunately, little work has been carried out on related deposits west of the divide except for the work by Harrison (1976) and Ferguson and Osborn (1981). Their work on the stratigraphy of Glacial Lake Elk, an ice-dammed lake created by ice in the Rocky Mountain Trench, showed that deglaciation of the area occurred before approximately 12,000 to 14,000 years ago and that ice probably has not advanced through the Elk Valley subsequently. Harrison also suggested (Harrison, 1976, p.170) that because the Kananaskis and Elk Rivers share a common divide, occupied by the Mount Joffre Ice Field, it is also unlikely that ice flowed down the lower Kananaskis Valley subsequent to this time. He did state, however, that this does not preclude smaller advances which failed to extend below the level of Glacial Lake Elk (elevation 1375 m) and cited evidence to suggest that such advances have probably occurred. The date he furnished from Glacial Lake Elk is still quite clearly a Late Wisconsin date albeit a minimum date for this event.

Luckman and Osborn (1979, p.57) described high elevation kames and eskers at one site in the Main Ranges of the Rocky Mountains. These reflect massive ice stagnation during the deglaciation of the Late Wisconsin advance. They also noted the presence of erratics of metamorphic material (Miette Formation) which suggested that the source of the last ice to deposit material in their study area originated from west of the Continental Divide, in the Interior of British Columbia. The ice-stagnation site studied by Luckman and Osborn (1979) is in an area adjacent to a contemporary glacier. Even a modest glacial advance should have removed these ice stagnation features.

Perhaps, at a general level, it is productive to compare the evidence discussed above with the proposed glacial chronologies for the Interior of British Columbia. A close correlation of events is not expected but because the two areas occupy opposite sides of the Eastern System of the Western Canadian Cordillera it is possible that there might well be some similarity in their respective records. Excellent summaries of the data from this area are found in Denton and Hughes (1981) and Wright (1983). Clague (1975a, 1975b) described evidence of a late-Wisconsin glacier which is thought to have flowed southeastward down the Rocky Mountain Trench into Montana. He associated upper ice limits as high as 2260 m with this advance. The southern margin of the Late-Wisconsin advance (Fraser Glaciation) is the least controversial and best documented margin of the Cordilleran ice sheet. According to Fulton and Smith (1978), Clague *et al.*, (1980), and others, it appears that the maximum extension of this advance occurred between about 17,000 and 18,000 years ago. A map (Figure 3.1) published by Waitt and Thorson (1983) summarizes the results of researchers from both sides of the international boundary. It shows that the ice surface to the west of the Continental Divide was above an elevation of at least 2000 m all across southern British Columbia except for the lower Fraser River valley. Deglaciation appears to have been rapid because Glacier Peak (G) tephra has been found interbedded with post-glacial sediments in the Purcell Trench near the International Border (Waitt and Thorson, in Wright, 1983). This implies that deglaciation was occurring in this area 12,750 +/- 350 years ago (Porter, 1978).

The studies discussed above generally lend support to the argument that the maximum ice limits during the Late Wisconsin advance were not restricted to major valleys, but rather extended a considerable distance beyond the mountain front. This

provides support for a simple model of the glacial history of the Rocky Mountain Foot-hills. Although more research is required before the conflict between this and other models can be resolved, it is worth remembering that it is inadvisable to develop a model which is more complicated than necessary. In the natural sciences we are often forced to employ inductive arguments based on empirical evidence as opposed to deductive arguments which are based on established laws and principles. Inductive arguments must necessarily be couched in probabilistic terms and the conclusions may never be stated with absolute certainty. This is particularly true for conclusions regarding past or future events. This implies that any explanation which employs all of the observational data and is not a tautology, or self-contradictory, has some probability of being correct but the probability is inversely related to the number of assumptions that are required by the explanation. Therefore, until there are observational data to the contrary, one should assume the simplest possible model to explain a set of observations.

Applied to this study, this means that the geomorphic and stratigraphic evidence for multiple glaciations ought not be assumed unless there are some observations that are not explicable under the premise of a single event. Closer adherance to this principle is not likely to resolve all of the problems facing the development of a regional glacial history for the eastern slopes of the Canadian Rocky Mountains, but it would certainly remove a great deal of the controversy which presently exists. Because the model proposed in this study is very simple, probably the most important contribution it can make lies not in the evidence which supports it but rather in the evidence which refutes it. The falsifying evidence, if produced, will generate improvements which will allow the development of a more powerful explanatory model.

6.5 CONCLUSIONS

From the field evidence illustrated on the Landforms Map it is clear that most of the study area reveals Quaternary landforms associated with active ice. Exceptions occur within the Rocky Mountain Foothills but not adjacent to the major valley systems. It is interesting that geomorphic processes of deglaciation, though subsequent to the formation of the active-ice landforms, did little to alter the landscape. Scattered erratics lowered onto the surface, crevasse fillings imprinted on the drumlins and flutings, the

esker complexes and many of the meltwater channels comprise the landforms associated with deglaciation. Nested end moraines, and nested lateral moraines were not identified. It would appear that either the glacier(s) were transporting little sediment at the time of deglaciation, or that the ice became stagnant and downwasted *in situ*. Meltwater channels to the east of the Foothills are mainly oriented towards the southeast. Large, clearly defined channels occur to the southeast of the study area and these carried meltwater from that portion of the study area which lies east of the Foothills. The major meltwater channels of the region are thus found beyond the study area, north of Rocky Mountain House, between the Nordegg and Baptiste Rivers, and to the east of the North Saskatchewan River. Boydell *et al.*, (1972) mapped many of these channels. The discontinuous nature of the meltwater systems may reflect either pronounced areal wastage of the related ice, so that drainage divides were very ephemeral, or rapid deglaciation whereby the channels were not occupied for sufficient duration to form well-integrated networks.

The occurrence of hummocky moraines along the eastern slopes of the Foothills attests to the relatively inactive ice mass which stagnated against these topographic barriers. The paucity of ice-marginal moraines in an area where ice of the Brazeau and Athabasca Lobes should have been subjected to longitudinal compression (in response to the increase in elevation of the underlying topography) is difficult to interpret but it is thought that ice from the Brazeau Lobe expanded eastward from the Foothills and forced the Athabasca Lobe away from the eastern slopes, thus precluding the formation of lateral moraines.

The landform assemblages, and consequently the proposed model of deglaciation, are significantly different from those found along the eastern slopes of the Rocky Mountains further south where well-preserved outwash terraces and meltwater channel systems grade into nested moraines but subglacial active-ice landforms are rare.

6.6 FUTURE RESEARCH

The difference between the glacial history of the study area, as interpreted from geomorphic evidence, and that provided by glacial stratigraphers in the area (Reimchen and Bayrock, 1977) and Roed (1968, 1975) and Boydell (1972, 1978) in adjacent areas, underscores the need to develop a more comprehensive model drawing on the expertise of several of the earth science disciplines. A broader data base would identify areas where the evidence produces inconsistencies and additional tests could be generated in these areas to resolve some of the problems prior to publication of the results. Within the Brazeau River valley there is still much useful research left to do:

(1) The hummocky moraine (lateral moraine ?) north of Thunder Lake at the confluence of the Brazeau and Crooked Creek valleys should be studied in detail.

(2) The moraines near the headwaters of Ruby Creek should be traced to attempt to locate end positions.

(3) The lake core evidence from Fairfax Lake should be replicated and Muskiki Lake should be re-cored in several locations to verify, or modify, the dates obtained in this study.

(4) The geomorphology of the area in the vicinity of the Brazeau Reservoir should be studied in order to provide additional information about the contact zone between Laurentide and Cordilleran ice.

Considerably more work is required along the full extent of the eastern slopes of the Rocky Mountains. Perhaps the recent archeological studies in the Bow River valley will stimulate interest in the area. White, J. of Parks Canada is quoted in the May 3, 1984 issue of the *Edmonton Journal* as stating that the archaeological work associated with the widening of the Trans-Canada Highway near Vermilion Lakes has unearthed bones, charcoal and projectile points from an early human habitation site. Radiometric dates have apparently been obtained which indicate an age of between 11,700 and 10,900 years B.P. (D. Cooper, Edmonton Journal, May 3, 1984).

Roberts, A. (Simon Fraser University) (pers. comm.) has expressed interest in using sophisticated remote sensing techniques to detect thin surficial deposits such as beach deposits and the boundaries between dissimilar unconsolidated deposits. It is thought that perhaps such a study may furnish valuable data for locating relict ice-marginal lake

margins (and potential archeological sites) and glacial limits. It is probable that the renewed interest in the Foothills and the Front Ranges of the Rocky Mountains will see a significant increase in the amount of information available pertaining to the glacial history of this part of Alberta.

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APPENDIX ONE

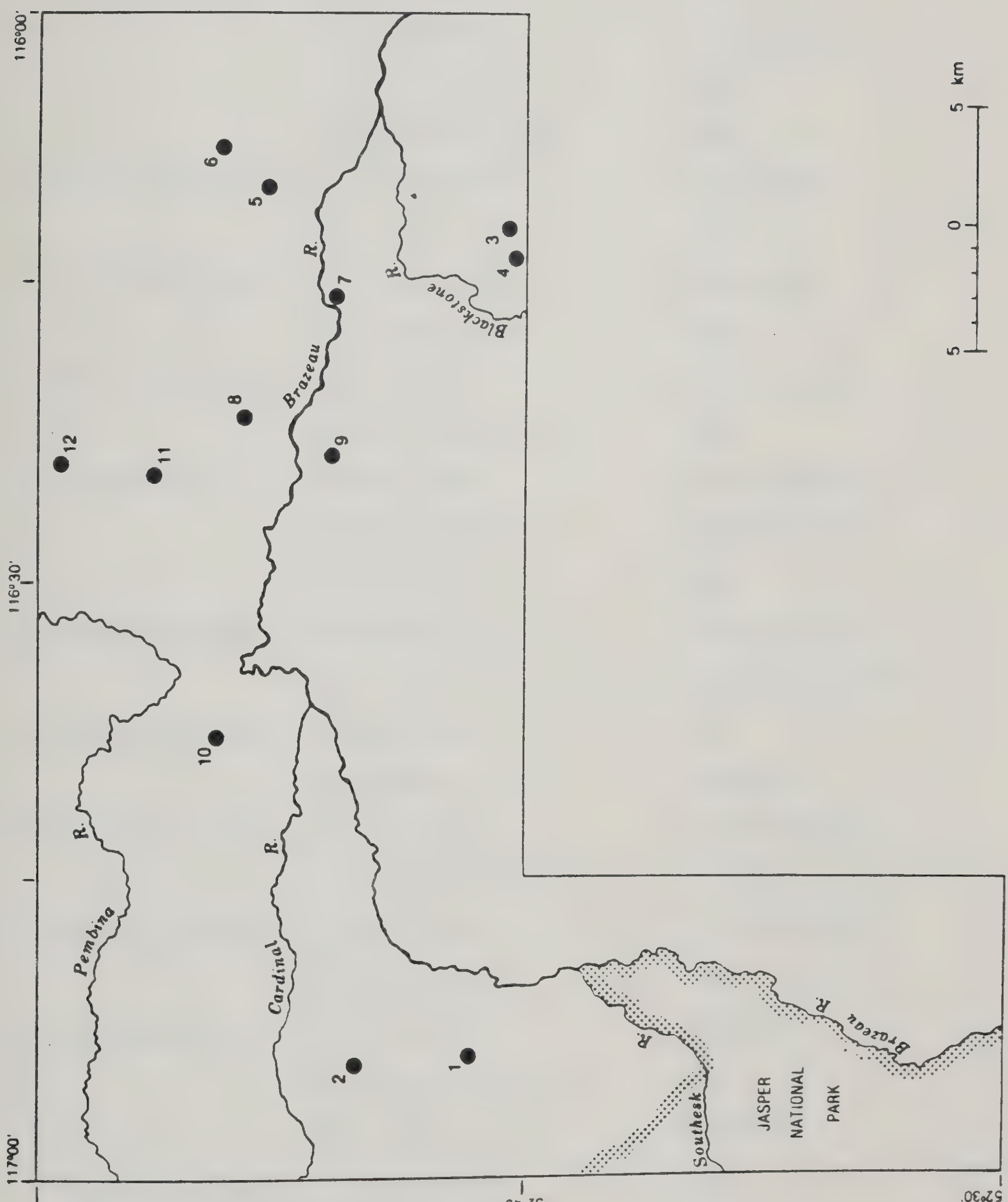
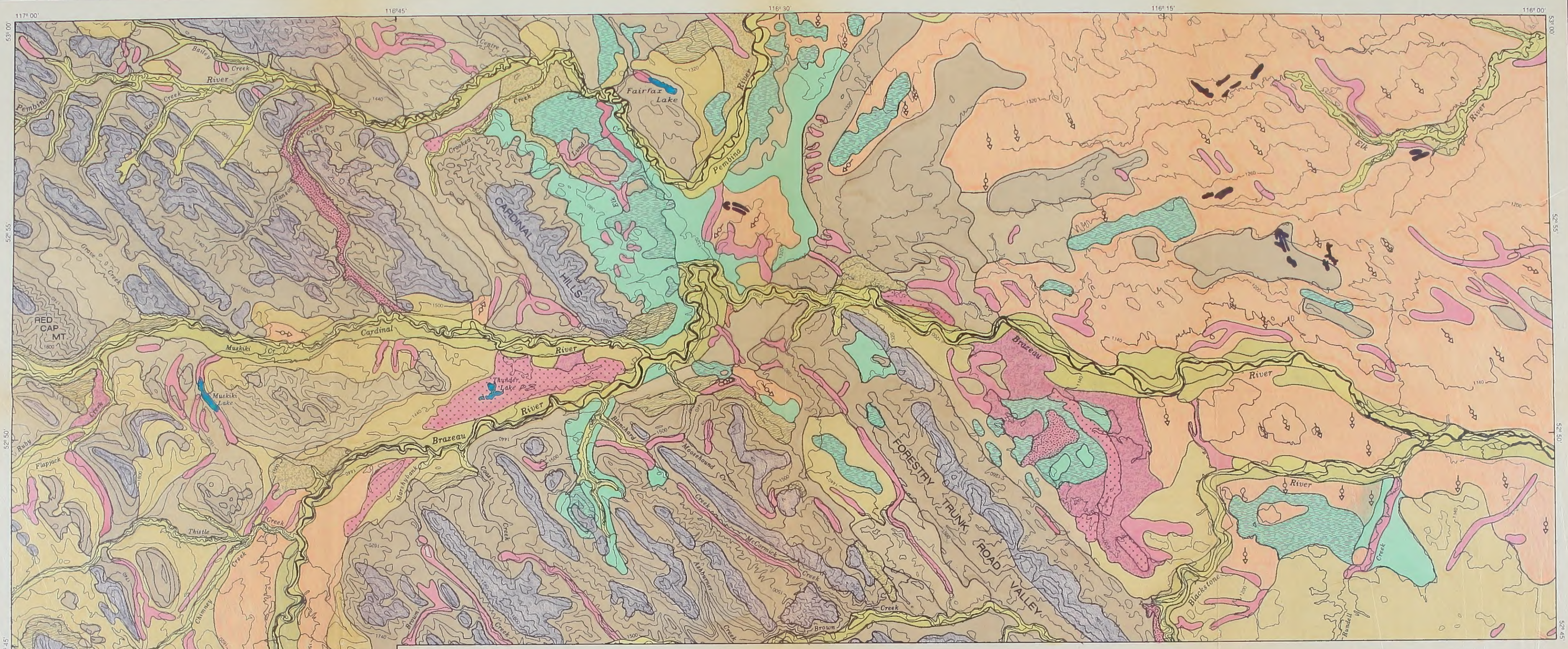
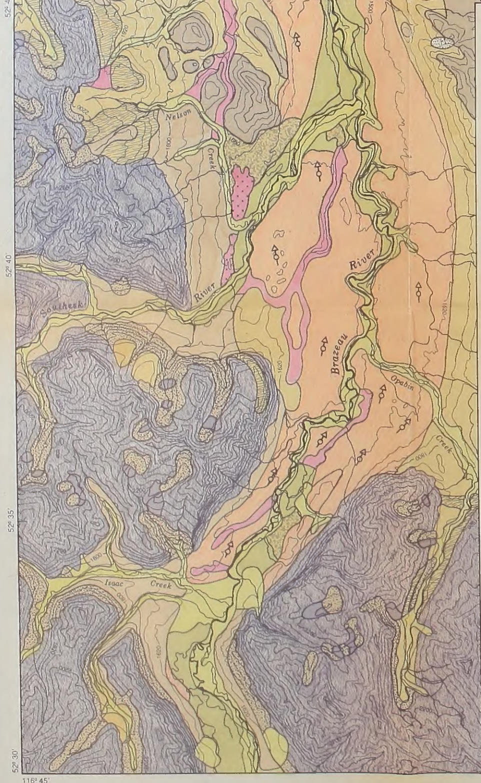


Fig. A1.1 Research Council of Alberta Drill Logs

SELECTED WELL LOGS: IMPERIAL OIL MUSKIKI PROJECT, 1976

LOCATION	DEPTH	MATERIAL
1=SW17/44/20/W5	0′– 2′ (0–0.6 m)	clay
	2′–14′ (0.6–4 m)	clay and gravel
	14′–40′ (4–12 m)	gravel
	40′–400′ (12–122 m)	shale
2=NW31/44/20/W5	0′–25′ (0–7.5 m)	unspecified
	25′– ? (7.5 m– ?)	shale
3=SW1/44/16/W5	0′–19′ (0–5.8 m)	brown clay
	19′– ? (5.8 m– ?)	shale
4=SE1/44/16/W5	0′–20′ (0–6 m)	sand and gravel
	20′–70′ (6–21.4 m)	shale
5= 20/45/15/W5	0′–4′ (0–1.2 m)	clay and boulders
	4′–12′ (1.2–3.7 m)	weathered shale
	12′– ? (3.7 m– ?)	shale
6=NW28/45/15/W5	0′–45′ (o–13.7 m)	clay and boulders
	45′– ? (13.7 m– ?)	unspecified bedrock
7=NW3/45/16/W5	0′–6′ (0–1.8 m)	clay
	6′– ? (1.8 m– ?)	sandstone
8=NE24/45/17/W5	0′–42′ (0–13 m)	clay and boulders
	42′–50′ (13–15.3m)	sandstone
9=NE31/45/17/W5	0′–10′ (0–3 m)	clay and boulders
	10′–21′ (3–6.5 m)	shale
10=NE24/45/19/W5	0′–10′ (0–3 m)	sandy clay
	10′–50′ (3–15.3 m)	clay
	50′– ? (15.3 m– ?)	sandstone
11=SW3/46/17/W5	0′–18′ (0–5.5 m)	sand
	18′–21′ (5.5–6.4 m)	gravel
	21′–80′ (6.4–24.4 m)	sandstone

12=SW27/46/17/W5	0'–10' (0–3 m)	sand
	10'–20' (3–6 m)	gravel
	20'–50' (6–15.3 m)	sandstone

I. PRE-PLEISTOCENE

BEDROCK UPLANDS

All bedrock uplands have been classified as a single unit regardless of age or composition. Invariably, bedrock uplands form the largest positive relief features and all are comprised of consolidated rock with a partial veneer of talus and colluvium.

II. PLEISTOCENE LANDFORMS

CREVASSE FILLINGS

Crevasse fillings are small (2 - 5 m high), well-defined ridges of till which form a trellis pattern. Generally the orientation of the major ridges is parallel or sub-parallel to local ice-moulded landforms. Minor ridges lie approximately normal to the major ridges. Commonly crevasse fillings are draped over other landforms with no apparent alteration of the underlying features.

PITTED MORAINE

This unit is similar to kame-and-kettle topography except that the sediments are predominantly till. The surface topography consists of irregularly-shaped closed depressions and mounds.

HUMMOCKY MORAINE

Hummocky moraine is characterized by randomly-oriented mounds and depressions. The sediments are dominantly till with minor inclusions of glacio-fluvial and glaciolacustrine material. Unlike pitted moraine surfaces, there are few abrupt breaks of slope and the topographic variability is more subdued.

LATERAL MORAINE

Lateral moraines are ridges, often asymmetric, comprised of till deposited at the former margins of active ice. Ice-proximal slopes may be as steep as 34° while distal slopes are usually in the range of 10° - 15°. Generally the morphology of lateral moraines has been poorly preserved and only in a few locations can they be mapped with a high degree of certainty.

ICE-MOULDED MORAINE

Drumlins and flutings were combined into a single map unit. The longitudinal axes of these features indicate the flow directions of the most recent active ice to affect the area. Generally the relative relief of ice-moulded ridges ranges from 2 to 20 m but maximum heights exceed 40 m. Till, similar in composition to that of adjacent ground moraine, is the most common constituent of ice-moulded ridges and no rock cored drumlins or flutings were observed within the area. Irregular transverse ridges with drumlinoid mounds, horns or tails extending down valley are included in this unit. These are considered the equivalent of crested drumlins described by Lundqvist (1970).

GROUND MORAINE

Ground moraine consists of extensive till sheets which mask and subdue the relief of the underlying bedrock. The sedimentary characteristics of the till are highly variable. Where exposed the till sheets commonly range between 2 and 5 m in thickness.

MORAINE-MANTLED BEDROCK

This is a gradational unit between bedrock uplands and ground moraine. In most locations a large amount of local colluvium has been incorporated into the till. The thickness of the till veneer is highly variable, subduing the local bedrock topography, but not masking it.

LACUSTRINE-MANTLED GROUND MORAINE

Ice-marginal ponding during deglaciation resulted in a mantle of glacial lake sediments superimposed on the underlying landforms. Usually this mantle is thin, often discontinuous and the topography of the underlying deposits is still apparent. Because these are necessarily areas of subdued topography and poor surface drainage, organic deposits mask much of the surface.Up to 6 m of lacustrine material overlies till in some locations, but commonly the mantle consists of sand and silt about 1 m thick.

MELTWATER CHANNELS

Classic examples of ice-marginal and proglacial channels may be identified but they have been mapped as a single category. Marsh, sloughs and misfit streams usually occupy the valleys, making them relatively conspicuous features. Frequently the channels are not well-developed fluvial valleys but, rather, are discontinuous channel segments. The valleys are generally steep-sided and narrow. Rarely are relict fluvial sediments observed along the valley floors and recent streams are most commonly incising into bedrock or colluvium.

OUTWASH TRAINS / PLAINS

Many of the outwash trains have been dissected by steam erosion and thus were mapped as outwash terraces. Other outwash surfaces form trains or plains encroached on in places by postglacial colluvium. Rare exposures of the sediments revealed coarse textured, rounded gravels.

PITTED OUTWASH TRAINS

These have sedimentary characteristics and surface configurations similar to outwash trains except that the surfaces display occasional kettles. Sands and fine gravels are the major constituents of these landforms. The large diameter, shallow depth, and irregular shape of the kettles suggest that these deposits were partly superimposed on stagnant ice.

OUTWASH TERRACES

Paired and non-paired outwash terraces flank the main rivers but are discontinuous surfaces not linked with obvious end moraine remnants. The sediments are generally coarse textured gravels, moderately to poorly sorted. Relict channel scars frequently mark the treads.

ESKERS

These steep-sided gravel ridges, usually flanked by kettle depressions, occur singly or in complexes. The larger eskers range from 3 km to more than 14 km in length. Some of the ridges are veneered by a thin layer of till.

GLACIAL LAKE BASINS

Proglacial and ice-marginal lake deposits are mapped as glacial lake basins where the lacustrine deposits are sufficiently thick to provide an obvious alteration of the landscape. In areas where lacustrine deposits mantle but do not significantly mask pre-existing relief the landform was mapped as lacustrine-mantled ground moraine. Numerous small lacustrine deposits were identified but they were too small or thin to be mapped at this scale.

III. RECENT LANDFORMS

STREAMS

Major stream channels, floodplains, bars, alluvial fans and low terraces, which are inset within older landform units, comprise this unit.

SCREE APRONS AND SCREE FANS

These are steeply sloping accumulations of colluvium, usually located immediately adjacent to bedrock uplands, and they often mantle Pleistocene landforms. Rock glaciers, because of their rare occurrences, are included in this unit.

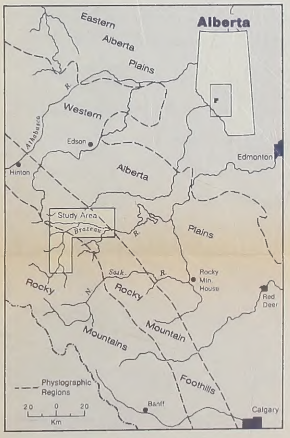
LANDSLIDES

Depositional lobes and associated erosion scars of major slope failures are mapped as landslides. These deposits invariably include some very large fragments and are generally the coarsest-textured sediments mapped.

MORAINES

Nested end moraines found near the upper ends of glaciated valleys, and in cirques, were interpreted as being younger (probably Holocene) than the Pleistocene landforms. They display fresh, or well-preserved, morphologies and are generally located within 2 km of the valley heads.

Ice flow direction inferred from selected drumlins.



Study Area Location Map

3,000 0 3,000
Metres

Contour Intervals, 60 m below 1,980 m and 120m above 2080 m

Topographic information from: Energy Mines and Resources, Canada: maps 83C - (10, 15 and 16)

Drawn by K. Davis and Associates, 1980

Geomorphology by D. R. Kvill, 1977-1979

Glacial Landforms of the Brazeau River Valley, Foothills of Alberta